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PHYSICS OF THE EARTH

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PHYSICS OF THE EARTH—VII

Internal Constitution of the Earth

EDITED BY BENO GUTENBERG
Professor of Geophysics, California Institute of Technology

Contributors

L. H. ADAMS
Geophysical Laboratory, Carnegie Institution of Washington

REGINALD A. DALY
Harvard University

B. GUTENBERG
California Institute of Technology

HAROLD JEFFREYS
St. John's College, Cambridge, England

WALTER D. LAMBERT
U. S. Coast and Geodetic Survey

JAMES B. MACELWANE, S.J.
St. Louis University

C. F. RICHTER
California Institute of Technology

C. E. VAN ORSTRAND
U. S. Geological Survey

H. S. WASHINGTON
Geophysical Laboratory, Carnegie Institution of Washington

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FOREWORD

It is generally agreed that more attention should be given to research in the middle ground between the sciences. Geophysics—the study by physical methods of the planet on which we live—is a conspicuous instance of such a middle-ground science, since it shades off imperceptibly in one direction or another into the fields of physics, astronomy, geology, to say nothing of biology, with which the subject of oceanography is closely connected. Some branches of geophysics, such as meteorology, terrestrial magnetism, geodesy and oceanography have long had a more or less independent existence, but it has become increasingly clear that these subjects, and many others, are all parts of geophysics. For various reasons, among which may be mentioned the development of geophysical methods in prospecting for oil and minerals, there has lately been a considerable development of interest in geophysics, but this development has not been matched by the publication in English of systematic treatises on the subject. With these ideas in mind, Dr. J. S. Ames, during his term as Chairman of the Division of Physical Sciences of the National Research Council, was instrumental in organizing in 1926 a large committee to prepare a series of bulletins on The Physics of the Earth, the purpose being "to give to the reader, presumably a scientist but not a specialist in the subject, an idea of its present status together with a forward-looking summary of its outstanding problems."

In due course subcommittees were formed to prepare reports on the following subjects:

- The Figure of the Earth
  - Gravity, Deflection of the Vertical and Isostasy
  - Tides, Ocean, and Earth
  - Variation of Latitude
- Seismology
- Terrestrial Magnetism
- The Age of the Earth
- Field Methods for Detecting Unhomogeneities in the Earth's Crust
- Internal Constitution of the Earth
- Meteorology
- Oceanography
- Volcanology

That this project, as ambitious as it is important, is now coming to fruition with the publication of these bulletins is due partly to the skill and farsightedness with which Dr. Ames selected the committee and assisted in outlining its program; partly to the care and interest with which Dr. Ames's successor, Professor Dayton C. Miller, directed the committee's activities during his term as Chairman of the Division; and particularly to the devotion with which the chairmen and members of the several subcommittees have carried out their respective assignments. The hearty thanks of the National Research Council and of the readers of these bulletins are due the several authors for their efforts.

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PREFACE

The Committee on the Interior of the Earth was appointed in 1926. In the following years, H. Jeffreys, R. A. Daly, J. B. Macelwane and H. S. Washington prepared manuscripts for a bulletin. However, owing to various circumstances, the work of the committee was delayed. In June, 1937, B. Gutenberg was appointed chairman of the committee with the task of reorganizing it and preparing the bulletin. The new committee consisted of L. H. Adams, Geophysical Laboratory, Washington, D. C.; F. Birch, Gordon Mackay Laboratory, Cambridge, Mass.; H. Jeffreys, St. John's College, Cambridge, England; W. D. Lambert, U. S. Coast and Geodetic Survey; J. B. Macelwane, S. J., St. Louis University; C. F. Richter, California Institute of Technology; C. E. Van Orstrand, U. S. Geological Survey; and B. Gutenberg, California Institute of Technology, chairman. The available manuscripts were revised by their authors, that of the late H. S. Washington being added to the bulletin after revision by L. H. Adams.

The difficulties involved in the writing of a book by several authors are well known. As it was impossible to hold a meeting of the committee, the only way to avoid too much overlapping of parts, serious contradictions of opinions and omission of important sections was an extended exchange of ideas through written communications among the collaborators dealing with related subjects. There was, however, partial exchange of manuscripts before they were printed, together with verbal discussion among some authors, so far as this was possible.

The attentive reader will observe that some differences of opinion and interpretation still exist among several of the collaborators. An earnest effort has been made to remove all mere misunderstandings and to clear up points of difference that turn out to be matters of terminology. There remain only those irreducible disagreements which are inevitable with reference to a subject that combines the results of so many divergent and highly active fields of investigation. Such disagreements are based largely upon differences in judgment as to the comparative validity of distinct kinds of evidence. Instead of glossing over them, the committee has tried to bring these points out plainly,

so that the critical reader may have as much assistance as possible in forming his own opinions.

The chairman is responsible for the final decision on the material to be included in the present volume, but each contributor has been allowed full expression of his own point of view in the sections allotted him.

B. GUTENBERG.

PASADENA, CALIF.,
May, 1939.

INTERNAL CONSTITUTION OF THE EARTH

CHAPTER I

INTRODUCTION

B. Gutenberg

Professor of Geophysics, California Institute of Technology, Pasadena, Calif.

The interior of the earth has invited speculation since the time when people began to think about things other than their own personal needs. The first scientific hypotheses were probably based on the fact that molten lava erupts from volcanoes, which suggested the idea that the interior of the earth is red-hot and molten. This idea seemed to be proved when over 100 years ago measurements indicated an increase of temperature of about 1°C. per 30 m. of depth (a critical review has been given by Thiene[1]). Doubts as to the validity of extrapolation from the data observed near the surface to a large depth have been expressed, *e.g.*, by Poisson, about 100 years ago, who did not believe in the possibility of a gaseous central part of the earth with a temperature of the order of hundreds of thousands of degrees, and by W. Hopkins (Philos. Trans., 1839, 1840, 1842), who thought (erroneously) that the observations on precession and nutation at least required a solid mantle for their explanation. He suggested—and perhaps was first to do so—that the lava of volcanoes ascends from local cavities near the surface filled with molten material. Lord Kelvin concluded from observations on the tides that the earth as a whole must be more rigid than glass (Philos. Trans., 1863).

Since that time, methods and results have been refined, and during the last 30 years data on the elastic constants have been added from the observation of earthquake waves. The first useful calculation of Poisson's ratio as a function of depth for the outer part of the mantle seems to have been made by Zoeppritz and Geiger in 1909, and the first useful calculation of the elastic constants as a function of depth by B. Gutenberg in 1923. However, the problem of whether the core of the earth is "solid" or "fluid" is still open. Though most specialists in this field agree that all observations indicate a large drop in rigidity at the boundary of the core, no final conclusions have yet been reached as to whether the rigidity and the viscosity in the core are so small that we would call the material fluid under normal conditions or solid.

The problem is still more complicated by the different use of certain words by various writers. The expression "a material melts" means fundamentally the behavior that follows when the temperature is raised above the melting point. At the surface, if a material melts, it becomes fluid, and therefore, the word *molten* connotes "fluid"; it has been widely assumed that under all circumstances a material is fluid above its melting point (neglecting delay in melting). However, there are two different processes which are frequently confused, since they usually run parallel under the conditions at the surface of the earth: the transition from the crystalline to the noncrystalline state, and the transition from relatively large rigidity and viscosity to values close to zero.

The transition in the first case is clearly defined by the melting point, although the process is frequently delayed. It is marked by a discontinuity; either the molecules are arranged in a certain regular order—crystals—or they are not. Unfortunately, there is no word simply and solely describing the transition from the crystalline to the noncrystalline state.

The second type of transition is that from solid to fluid; it need not be connected with the first. Fluid is frequently thought of as meaning "having small rigidity" or "having no rigidity." However, there is no physical object that truly has zero rigidity, and there is no discontinuity between fluid and solid. Unfortunately, the determination of small rigidities in the laboratory is rendered impossible by the fact that materials with very small rigidity have, at the same time, a small viscosity, so that the effect of plastic flow is much larger than the effect of elastic shear in such objects. The smallest values of the coefficient of rigidity determined thus far are of the order of magnitude of 10^9 dynes per square centimeter and belong to rubber and similar substances. As the largest values found in laboratories for materials like steel or nickel are of the order of 10^{12} dynes per square centimeter, one gets the impression, from extrapolating, that a material with a rigidity of the order of 10^6 dynes per square centimeter or less would be considered fluid, in the common sense of the word.

Other authors define a liquid as having no *strength* (minimum stress that must be applied in a body to produce plastic *flow*). This definition would combine flow and fluid, but it would not follow the usual use of the word fluid, since then material with practically zero strength and high viscosity—as possibly in the deeper parts of the mantle of the earth—would through this definition be fluid, although it could take thousands of years until the movement would become traceable. Finally, materials with a very small coefficient of viscosity

rapidly assume the shape of any containing vessel. This procedure is considered the definition of a fluid by many.

A hard and fast definition can be made only by international agreement, and so long as such a one does not exist, it will be best to give numerical values of the coefficients involved where there is doubt. Temporarily, we shall use the word *fluid* to indicate a combination of relatively very small viscosity, strength and rigidity.

A solid noncrystalline material (with large rigidity) is called *vitreous* regardless of whether its temperature is above or below the melting point of the material. If its strength is small, it may undergo plastic flow in the course of time. The speed of this flow under a given stress difference (above strength) is determined by its viscosity. Thus, a solid material may flow in the course of time. Such behavior is occasionally indicated in the literature by addition of the word *secular* to a property, but this procedure will not be followed in this book. The problem of which parts of the earth are crystalline and which are vitreous is not yet solved.

More successful were the attempts to find the approximate density as a function of depth. Experiments to find the mean density of the earth were made by Maskelyne and Hutton in 1774 and by Cavendish in 1799. Many others improved upon the methods in the last half of the nineteenth century. Though at first it was generally assumed that the density increases gradually to a maximum in the center of the earth, Wiechert concluded in 1896 that a rocky mantle with a thickness of about 1,200 km. surrounds an iron core and that in each the density does not vary greatly. Though he drew this conclusion from theoretical reasoning on density and its effects, the development of seismology about 30 years ago provided further details concerning the interior of the earth. The first attempts by Milne and Benndorf between 1903 and 1906 to find the velocity of earthquake waves in the interior of the earth gave only the order of magnitude. The first good approximation for the mantle was published in 1907 by Wiechert and Zoeppritz. In 1906, Oldham found about 2,500 km. for the radius of the core. This was too small because he misinterpreted the observations for large distances; the radius of about 3,500 km. found by Gutenberg in 1913 is still considered correct.

The structure of the earth's crust was investigated later. The first important result from the study of earthquake waves was obtained by A. Mohorovičić in 1909 when he discovered the discontinuity forming the bottom of the continental layers. The depth of 60 km. that he found was later reduced to about 40 km. and less, depending upon the region. In 1921, Tams and Angenheister found inde-

pendently that the velocity of the surface waves is greater along ocean bottoms than over continental areas. Tams concluded that the sial, which forms the uppermost crust in the continents, is absent under the oceans. By means of more detailed observations and refined methods, Gutenberg, in 1924, concluded that this sialic part of the crust is missing only in the Pacific Basin but exists at the bottom of other oceans, although it is thinner there than in the continents.

The fact that two different types of properties, one concerning physical state, the other density and elasticity, are considered in describing certain parts of the earth again produces difficulties in the terminology. One of the words most frequently used in the literature of this field is *crust*. However, many authors use it to indicate that it is crystalline, in contrast to the vitreous *substratum* (Osmond Fisher). In a similar way, *lithosphere* indicates the shell with large strength overlying the *asthenosphere* with relatively small or practically zero strength (J. Barrell). *Sial* and *sima*, on the other hand, refer to composition; the first indicates *si*licates, predominantly of *al*uminum, sodium and potassium, and the second that of *si*licates, predominantly of *ma*gnesium and calcium and containing less aluminum, sodium and potassium than the sial (E. Suess). If none of the foregoing properties is to be stressed, neutral words, such as *layer*, *shell*, *sphere*, should be used. Thus, the outer parts of the earth in which relatively low velocities of earthquake waves prevail are called *continental layers*, as they are characteristic of the continents, although they also seem to form the upper part of the bottom of the Atlantic and Indian oceans. However, many of these regions at one time during geological history have been a part of a continent. The expression *sialic layers* is used frequently as synonymous with *continental layers;* however, the deepest parts of the continental layers may be simatic. That part of the earth between the surface and the discontinuity at a depth of about 2,900 km., where the elastic constants and probably the density change discontinuously, but excluding the continental layers, is called the *mantle*, and for the remaining part inside this discontinuity the word *core* is used.

From the theory of isostasy, another "discontinuity" has been inferred by many authors. In 1855 Pratt and Airy discovered independently that, to a first approximation, gravity is the same everywhere in a given latitude (for details see vol. II of this series). This fact leads to the conclusion that there exists a *depth of compensation*, at which there is approximately hydrostatic equilibrium. However, this depth is not a discontinuity, nor is it accurately defined; it depends on the deviations from hydrostatic equilibrium which are allowed in the

definition and which are not known as a function of depth. As hydrostatic equilibrium presupposes zero strength, the depth of compensation is for all practical purposes identical with the boundary between the lithosphere and the asthenosphere, which, again, is not a discontinuity but depends on definitions.

Though for the density and the elastic constants inside the earth (except for the rigidity in the core) values have been found that are accurate within the limits needed for general problems, many other quantities and properties concerning the earth's interior and its history are rather doubtful. This is true, for instance, for the temperature, the viscosity, strength, the forces producing changes, especially the formation of mountains, the question of movements of parts or the whole of the crust. Thus, in this book, the "outstanding problems" and results marked by "probably," "perhaps" or "possibly" or given numerically with the remark that they may be wrong by one order of magnitude or even more are quite numerous. The difficulty is partly due to the fact that the problems require cooperation of an unusually large number of sciences, partly to the circumstance that the chief science, geophysics, as an entity is very young and only recently has begun to find the support needed for its development. The earliest textbooks containing at least some data that are approximately correct on the interior of the earth are those of Schmidt,[2] A. v. Humboldt[3] and Naumann.[4] The progress made during the second half of the nineteenth century stands out clearly when these books are compared with those of Fisher,[5] Günther,[6] Arrhenius,[7] Rudzki[8] and especially Trabert.[9] Shortly after 1900, E. Wiechert started instruction in geophysics at the University of Göttingen, the first course in which practically all fields of geophysics and geodesy were covered in a modern way. Unfortunately, these lectures have not been printed.

The most rapid development of general geophysics, especially with regard to our knowledge of the interior of the earth, has taken place during the last 15 years, as may be seen from the references in this book. For this reason, it has seemed necessary again and again to summarize the results in textbooks. The facts that the rapid increase in research has resulted in specialization by the investigators and that more and more knowledge in other sciences has been required make it impossible for one writer to cover the whole field. Therefore, many of the textbooks published during recent years have been written by means of the collaboration of small groups of authors. The first instance of this kind is the series "Einführung in die Geophysik."[10] Other writers have restricted themselves to special fields. A review of the problems and results concerning the interior of the earth was

given by Gutenberg.[11] The fundamental book on the physical aspect is "The Earth" by Jeffreys.[12] Three textbooks[13,14,15] on geophysics, of a more general character and written by several authors in collaboration, give much information on the physics of the interior of the earth, and three others[16,17,18] place more stress upon the geological side of the problems involved. An attempt to provide a more detailed handbook concerning the problems of the physics of the earth, including the borderland problems, has been made by Gutenberg.[19] Here belong also the series "Physics of the Earth," of the National Research Council, which are listed in the Foreword of this volume.

Much information in regard to the interior of the earth is to be found in periodicals on general geophysics,[20] and the results of other investigations into such problems are scattered among a large number of periodicals in the fields of physics, geodesy, magnetism and electricity, geology, geography, seismology and other sciences and among the publications of various learned societies. The finding of references is aided by special periodicals giving abstracts.[21]

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CHAPTER II

THE ORIGIN OF THE SOLAR SYSTEM

Harold Jeffreys
St. John's College, Cambridge, England

1. The problem of the origin and development of the solar system suffers from the label "speculative." It is frequently said that as we were not there when the system was formed we cannot legitimately arrive at any idea of how it was formed. Similarly, the investigation of the nature of the earth's interior has been condemned because we cannot personally penetrate more than a limited distance into it. It is perhaps less frequently realized that such considerations, if accepted, would destroy not only cosmogony and geophysics but also most of our everyday knowledge. Our direct experience consists simply of sensations, and these, if they can be considered to have any spatial location at all, are within our own bodies; all external objects are not known directly, but are inferences from our sensations or concepts introduced to explain our sensations. The difference in directness of knowledge is one of degree, not of principle. In writing this article I am aware directly of certain visual, auditory and tactile sensations; the typewriter and the sheet of paper are inferred from them and I say that I see, hear or feel them. Another sound that I notice I attribute to a piano some staircases away; the piano and its player are inferred from the auditory sensation alone. I go downstairs and speak to somebody on the telephone; his existence again is inferred from the sound in the receiver, but everybody is apparently willing to accept the inference that the sound is made by variations of electric current in the telephone wire. The existence of the electric current is not known directly by any sensation; it is a concept introduced because it helps to coordinate a number of observable phenomena. This coordination is achieved by means of a number of physical laws established in the laboratory. The process of inference from sensations by the construction of concepts and laws relating them is a regular part of ordinary life.[27,28] When we light a fire to heat a room, we are relying on inferences from the previous behavior of certain materials; when we design a steam engine to draw a given load, we are doing the same; when we sow wheat and expect a crop to grow, we

are still drawing inferences from sensation by using the technique of forming concepts and laws.

The problems of geophysics involve no new principle. We investigate the properties of the interior of the earth by studying elastic waves through it, its gravitational attraction and its tidal phenomena. All of these depend on laws detectable in the laboratory; solids do transmit two kinds of elastic waves, and fluids one; heavy masses do disturb one another by gravitation; the laws of tidal motion are statements, adapted to the particular circumstances, of the laboratory determinations of the laws of motion of solids and fluids. In studying the evolution of the solar system, again, we proceed by inference, using the same laws, with the slight modification that the law of gravitation is found to hold much more accurately in planetary motions than can possibly be ascertained in the laboratory. We proceed, however, backward in time. In the practical problems of inference involved in ordinary experience, we usually go forward in time, but not always. Legal procedure is an obvious instance to the contrary, the problem being to ascertain from present evidence what happened on a previous occasion. There is therefore nothing new in principle in attempting to reconstruct the past from present data.

2. All cosmogonic theories have in common that they postulate a state of the solar system in the remote past very different from the present one. It may legitimately be asked, before we consider any theory at all, whether there is any reason to suppose that the system ever was so different. There are at least four such reasons: radioactive changes, tidal friction, internal friction in the matter producing the zodiacal light, and aggregation.

2.1. All rocks are known to contain small, but appreciable, amounts of the radioactive elements uranium and thorium, which are continually breaking up and emitting heat in the process. The only known way of disposing of this heat is by conduction to the surface of the earth, from which it is radiated into space with the much larger amount of heat received from the sun. Conduction takes place only toward the places of lower temperature, so that this outflow of heat requires a hot interior. It appears that the radioactivity of the crust alone is capable of maintaining a temperature of about 500°C. in the interior.[1] Now the radioactive elements decay slowly; the measured rates imply that, of a given specimen of uranium, 1 part in about 6,000 million breaks up every year and, of thorium, about 1 part in 20,000 million.[2] If we go back some tens of thousands of millions of years, therefore, we come to a time when the radioactive elements were much more abundant and able to maintain a temperature in the

interior equal to the melting point of any known rocks. We therefore infer a state where the whole of the earth, except a thin surface crust, was fluid; but seismological evidence shows that it is now solid halfway to the center.

2.2. Tidal motions in the sea are familiar to all; the distortion of the solid earth by the tidal action of the sun and moon is less familiar but can be measured. All motions of fluids over solid boundaries involve frictional forces, mechanical energy being converted into heat; and all varying distortions of solids involve similar losses of mechanical energy. Both types of tide therefore reduce mechanical energy. The interaction between the earth and moon, however, must be such as to give no change of their total angular momentum, and it is found that the only way of obtaining a loss of mechanical energy (kinetic and potential together) is from changes in the earth's rotation and the moon's orbital motion; the earth's rotation gets slower and slower, and the moon goes further and further off.[3] Allowance for the sun's tides and for those raised in the moon by the earth requires treatment in greater detail, but the general results are the same. It is found that they imply an observable phenomenon, the apparent secular acceleration of the moon's motion, the amount of which agrees with what is inferred from the facts of tidal motion. Now, if we proceed backward in time, we find the moon nearer and nearer the earth, and about 4,000 million years ago[1] it should have been so close as to revolve in about 5 hr., the earth rotating in the same period. Further, this state is unstable; if the moon had been nearer it would have revolved more rapidly than the earth rotated, and in these conditions it would have approached the earth and been reabsorbed in it. This suggests that either the moon, or the seas, or both, have existed for less than 4,000 million years.

2.3. The zodiacal light, a cone of faint luminosity seen after sunset and before sunrise, is generally believed by astronomers to be scattered from a sheet of gas surrounding the sun and lying near the ecliptic. It appears to be a gas and not solid particles from the nature of the polarization of the light. Gutenberg[4] quotes from Schmid an argument that the phenomenon is atmospheric and not cosmic, because its visibility is correlated with the state of twilight; but this is surely to be expected of any diffuse and faint object whose radiation has to penetrate our atmosphere. The question could perhaps be finally settled if the Doppler effect were observed. At any rate, if astronomers are right, such a gas must be in orbital motion around the sun and must be losing energy through viscosity. The result must be that the outer parts are driven off into space and the nearer absorbed

into the sun. A time scale for the changes has not been fixed, though it probably could be; but we can see qualitatively that the matter should formerly have been more abundant.

2.4. Aggregation is seen in the continual appearance of meteors in the atmosphere and their conversion into dust. This implies that the earth is continually sweeping up meteors in its orbital motion about the sun, and that their number is therefore presumably decreasing. It is possible that they and the zodiacal matter are being regenerated, perhaps from comets, but this possibility merely implies that the amount of cometary material is decreasing. In either case there is an irreversible change. The rate can be estimated roughly. The distance between the orbits of Venus and Mars is about 100 million kilometers; the earth's diameter is about 12,000 km. Suppose that the meteors capable of coming near the earth's orbit are extended through a distance of 200 million kilometers at right angles to the plane of the ecliptic. Now consider a section of the system by a plane through the sun perpendicular to the ecliptic. The earth crosses this plane once every year. The area of the section of the annulus is 2×10^{16} sq. km.; the earth's section has an area of about 10^8 sq. km. So 1 part in 200 million of the region is swept out every year. Any given meteor, we may say, is equally likely to be anywhere within the belt when the earth crosses it, whence we infer that one meteor in 200 million is picked up every year. We infer that a few hundred million years ago the total number of meteors was many times what it is now. The estimate is very rough and could be improved by a more accurate treatment of existing data. All we need at present, however, is to notice that here again, if we look back a sufficiently long time, we find a marked change in the state of the system.

2.5. The clearest evidence is given by radioactivity in another aspect. Uranium and thorium in their decay give a number of intermediate products and finally yield two forms of lead, which change no more. Now lead, like uranium and thorium, occurs in all rocks but in smaller quantities.* It was pointed out by H. N. Russell that this fact provides a way of fixing an upper limit to the age of the earth. For if the earth existed more than a certain time ago, the uranium and thorium would have produced more lead than actually exists, even if there was no lead there to begin with. If there was lead originally,

* This seems strange at first sight, since lead is usually considered a common metal and the radioactive elements rare ones. But the abundance of lead as we ordinarily meet it is due to its occurrence in the easily recognizable mineral galena, which represents only a small fraction of the lead actually in the crust but generally distributed.

the time is correspondingly shorter. It appears from the most recent investigations that if the earth existed 3,000 million years ago it could have contained no lead[5] and that we cannot go further back without involving ourselves in inconsistencies. Our problem is then to say what can have been the state of the system still earlier. It appears that there are only two alternatives: the earth must have acquired its radioactive elements from outside at some more recent date, a hypothesis that has never been seriously considered; or it must itself have come into existence as a separate body more recently.

It is interesting to notice that these various methods suggest time scales that are not very different, and even if one of them should be wrong, it is very unlikely that all are. Their implications can be escaped only by purely *ad hoc* hypotheses without experimental support; and they imply that at some time under 3,000 million years ago the solar system was very different from what it is now, the earth and presumably the other planets having only just come into existence. To understand our present data some theory of the origin of the system is therefore necessary.

It may be remarked here that application of the uranium-lead-ratio method to specific minerals has led to age-determinations up to 1,800 million years, which is therefore a lower limit to the age of the earth. The age of the earth is known within a factor of 2.

2.6. A similar time scale is suggested by the phenomenon of the *expanding universe.* The observed fact is that the distant nebulae and clusters show systematic displacements of their spectral lines to the red, which would naturally be interpreted as being due to a velocity of recession increasing with the distance. The velocities reach about 20,000 km. per second. This changes all the wave lengths of the light reaching us by about 1 part in 15; the effect is indeed so great that if we had only the near and distant spectra, without the intermediate ones that provided the connecting links, we should hesitate to accept the lines as corresponding to the same elements. The data can be represented by an increase of the velocity of recession by 500 km. per second for every million parsecs (3.1×10^{19} km.). If the nebulae have maintained this velocity (and there is nothing in classical dynamics to alter it), they would all have been close together 6×10^{16} sec. or 2,000 million years ago. This can be interpreted as the age of the universe. Several different interpretations have been given by the workers on relativity. It is possible, for instance, to choose a unit of distance that will make the dimensions of the universe in terms of it constant with time, but apparently at the cost of making the lengths of our actual solid measuring scales decrease with time. The theo-

retical aspect of the matter appears to be very confused; all the workers on it appear to agree that results can be derived from premises that are intuitively obvious, but no two seem to agree with each other's premises or conclusions. Nevertheless, Hubble's spectral shifts constitute a very solid piece of observational evidence, and it is interesting to notice that the time scale suggested is of the same order as the age of the earth.

3. Scientific cosmogony may be said to have started with Buffon, who pointed out, in his "Natural History," that a number of observed similarities within the solar system can be explained by a single hypothesis. The observed facts were that (1) all the known planets (then six) move in nearly the same plane, (2) they all revolve round the sun in the same direction, (3) their rotations are in the same direction, (4) so is that of the sun and (5) so far as was then known the same applied to the revolutions of the satellites. His suggestion was that a wandering body of considerable mass, which he called a comet, struck the sun and drove from it a torrent of matter, which then condensed into sections to form the planets. Buffon's choice of the word *comet* seems to have been due more to lack of information about the nature of comets than to any doubt in his mind about the nature of the body required. On this hypothesis the similar sense of the revolutions and rotations would be explained as representing the plane of the original motion of the second body relative to the sun. All matter carried off would move in this plane or near it, and later changes would not alter this feature; and the rotation of the sun can be attributed partly to the shearing stresses developed during the collision and partly to the angular momentum of matter reabsorbed into the sun.

3.1. Laplace, in his "Système du Monde," objected to Buffon's theory that the original orbits would be highly eccentric, whereas the present ones are nearly circular. There are two possible answers to this objection. One is that the ejected matter would not necessarily be entirely collected into the planets; much of it would be spread through the system, and the planets would have to revolve through it. One effect of its resistance to the varying motion of the planets would be to reduce their eccentricities. There are two possible hypotheses about this medium: it might consist of solid particles (according to Proctor,[6] Bickerton[7] and Chamberlin and Moulton[8]) or a gas (according to Jeans and myself).

3.2. An alternative has recently been suggested by E. W. Brown. If a system consists of only two particles, they can revolve around their common center of mass indefinitely, the major axis and eccentricity of the relative orbit never changing. But if there are more

than two bodies, they all influence one another. The mutual potential contains harmonic terms whose arguments are linear combinations, with integral coefficients, of the longitudes of the smaller bodies with respect to the largest. These produce periodic disturbances of the elements of the orbits, which merely oscillate about mean values in periods determined by the mean angular velocities about the sun. There are other nonharmonic terms, but it is found that these do not affect the major axes of the orbits. They do, however, produce steady changes in the other elements, notably the eccentricities and inclinations of the orbital planes; these are technically called *secular*, somewhat unfortunately. For a closer analysis shows that these secular terms are not really so but represent vibrations of very long periods, running into millions of years. They differ from the first class of terms in that their periods depend on the masses of the planets as well as on their periods of revolution. It appears that the eccentricities of the planets' orbits cannot change greatly and have always been small. But it is found that if we try to obtain still more accurate solutions the mathematics becomes more and more intractable and that our solution represents only the earlier terms of series that ultimately diverge if the time interval considered is too great. If it reaches about 10^7 years, the solution fails.[9] It has been generally believed that the difficulty is purely mathematical and that, in spite of the difficulty of proving that they do not, the major axes and eccentricities cannot change greatly; but Brown suggests that this is not true. There are analogies to his suggestion in other branches of physics. A plane wave of sound, or a wave from a point source, can spread out indefinitely without loss of energy if we treat the gas as a continuous medium and ignore the effects of molecular structure. But as soon as we allow for these, we find that, as the wave advances, its regular motion is gradually transformed into irregular motions of the molecules and reappears as heat. An elastic wave in a crystal lattice with a regular pattern can theoretically travel similarly an indefinite distance; but as soon as we allow for irregular distortions of the lattice due to thermal agitation we find that, owing to the molecules being out of their equilibrium positions when the wave arrives, part of the energy is scattered or reflected. Thus there is continual loss of energy as the wave advances, and the scattered portion is converted into irregular short-wave motions, *i.e.*, into heat. The wave sent out when a stone is thrown into a river is another example. It initially spreads out symmetrically, and straight banks give regular reflections. But every irregularity of the bank produces its own reflection, these secondary movements reach other points in all sorts of phases and the result is an

irregular movement. Now in the solar system we may reasonably regard the varying masses and periods as irregularities and the variations of the velocities arising from the eccentricities and inclinations as irregular motions analogous to the thermal vibrations in crystals. If this is legitimate, we cannot expect the energy associated with any mode to remain permanently in that mode; there will be a continual interchange. Indeed, if the solar system were a set of particles vibrating about positions of equilibrium under some law of force, as in a crystal, instead of being in permanent revolution, this would be taken for granted. The ultimate result would be equipartition of energy, as in a gas. But there are two dubious points: Would the major axes as well as the eccentricities be affected in the long run by the interchange? And if the system is started arbitrarily, how long would it take to reach approximate equipartition? Mere interchange of eccentricities and inclinations would not reduce all of them, so that if we require a general reduction we must also appeal to changes of the major axes.

4. According to Laplace the solar system had one parent instead of two. It was a nebula extending beyond the present orbit of the most distant planet and revolving about the center. It condensed, as it lost heat, and rotated faster, having to keep the same angular momentum; at a certain stage the rotation became faster than the mutual gravitation of the parts could control. Then Laplace supposed that a ring was shed and that by some not very clear process this proceeded to collect into a planet. The process was repeated again and again; further, the planets repeated it and produced the satellites. The theory explained apparently the same facts as Buffon's did and also the smallness of the eccentricities, since each part of the nebula would be originally in nearly circular motion.

It must be said that Laplace stated his theory rather vaguely in a semipopular work and never developed it in detail. Its apparent plausibility at the time rested on two analogies, which are now known to be certainly false. Nebulae were known in his time, but their sizes were not. The typical nebula has a mass of millions of suns, and there is no observational ground for believing in the existence of a nebula anywhere near so small as our solar system. The other was the existence of Saturn's rings, which looked as if they might be a Laplacian ring that might ultimately become a satellite. Maxwell's work on rings of satellites, however, showed that they are stable in the condition required by Laplace; they would remain rings and could not collect into one body. Saturn's rings may be a broken-up satellite, but no planet or satellite is an aggregated ring.

The analogies, however, are less important than the question whether the theory explains what it claims to explain. It is not sufficient to point to various possible factors and to notice that they may in suitable circumstances produce effects in the right direction. In the early stages of a subject, we need first of all suggestive ideas that may serve as a basis for investigation. Buffon and Laplace both supplied these. But the further investigation must go into greater detail and consider in what circumstances, if any, the suggested processes would actually take place; whether the theory is internally consistent, *i.e.*, whether one part of it does or does not contradict another part; whether its conclusions agree with the facts where comparison is possible. Now Laplace's theory has turned out to be inapplicable to the solar system. Given the present distribution of mass and angular velocity in the system, we can calculate how fast it would have rotated at any given degree of distension; the more closely it is concentrated to the center, the faster it rotates. But the condition for the loss of an outer ring, or indeed of any portion of the outer material, determines the rate of rotation at the time. It is found that it could take place only if all the mass but about $\frac{1}{600}$ were in the central body; otherwise the moment of inertia would be too great for the required speed of rotation to be possible. This was first noticed by Roche. But Jeans[10] showed in 1917 that the ejected matter could condense only under certain conditions. Its density would have to be enough for its gravitational attraction to overcome the local tendency to spread through rotation, and this condition was found to imply that the density near the outside had to be comparable with the mean density of the whole. This contradicts the hypothesis of a strong condensation to the center and shows that the genesis of the solar system from a symmetrical nebula is impossible. Another variant of the theory, tested by the present writer, was found to lead to the same inconsistency.[1] Jeans finds, however, that agreement can be achieved if the angular momentum is greater, and constructs on these lines a theory of the formation of spiral nebulae and how stellar condensations form in their arms. Laplace's theory has nothing to do with the solar system as we know it, but it may have something to do with the formation of the sun in the first place.

The meteoric hypotheses of Lockyer and du Ligondès are open to the same tests as that of Laplace and fail for the same reason.

5. It appears that, if the planets had a stellar origin, we must consider them in any case to have been gaseous when they were first formed. This at once provides a criterion that must be satisfied by

any admissible theory. We know that the smaller bodies of the solar system, Mercury, the moon and the other satellites, are devoid of atmospheres, and an explanation of this fact is provided by the kinetic theory of gases. The molecules of all gases are in motion, the mean velocity at ordinary temperatures being of the order of 1 km. per second. It is in fact comparable with the velocity of sound. But a certain fraction of the molecules have velocities several times the mean. Now there is a definite limit to the velocity of a molecule that can be retained by a planet. If m is the mass of the planet, a its radius and f the constant of gravitation, the velocity cannot exceed $(2fm/a)^{\frac{1}{2}}$. If a molecule near the outside of an atmosphere exceeds this velocity, it will escape and afterward revolve round the sun. So planetary atmospheres are continually being lost. It appears that the loss is so slow as to be insignificant for the larger planets. The earth may lose a moderate fraction of its hydrogen and helium in 10^9 years if, as is now known from the study of meteors and the audibility of distant explosions, the temperature rises at great heights; but the moon and Mercury could not have retained oxygen and nitrogen.[11] A gaseous planet is all atmosphere. If it were too hot, it could not retain any of its constituents, but would spread out and cease to be a planet. Now when we apply this to any stellar theory of the origin of the planets, we are immediately faced with a problem. If the planet was ever gaseous, it would be both hotter and more distended than it is now. On the former ground, the molecular velocities, for the same substance, would be greater; on the latter, the velocity that the gravitation could control would be smaller. The boiling points of rock materials are somewhat vaguely known, but it seems that they become wholly gaseous at some temperature between 2,000 and 3,000°C. For a planet of known mass we can therefore estimate what its radius could have been in the gaseous state. It turns out that, below a certain mass, that radius is less than the radius in the solid state, which is impossible. For a material with the density of an average rock, the critical mass corresponds to a radius in the solid state of about 2,000 km.,[1] which is more than those of the asteroids and most of the satellites. Any smaller mass, if liberated as a gas into a vacuum, would spread out with a velocity comparable with that of sound and dissipate itself. There are, therefore, bodies in the solar system that cannot possibly have been formed by slow condensation from the gaseous state. No theory of gradual evolution can meet this criterion; these bodies, and therefore presumably all the others, must have been formed catastrophically.

[1] Ref. 1, p. 28.

5.1. We have a further criterion, based on the estimate of the maximum age of the earth. The sun is continually radiating energy, which implies, by the theory of relativity, a corresponding loss of mass. The rate of loss is 1.2×10^{20} g. per year. This looks a great deal, but when we estimate the loss in 10^{10} years, which is beyond any possible value of the age of the earth, it is still only about $\frac{1}{2000}$ of the mass of the sun. It appears therefore that when the system was formed the sun had practically its present mass.[12] Now, as Eddington once pointed out to me, it is a fact of observation that for stars of small mass, such as the sun, spectral type is closely correlated with luminosity, which is in turn closely correlated with mass; but spectral type and luminosity together determine the radius. If then the mass was nearly the same as at present, the same applies to the radius. This argument rests on observed astronomical facts and not on particular theories of stellar constitution, about which there is considerable room for disagreement; any theory that agrees with the facts would lead to the same result. The sun was in nearly its present state when the planets were formed. We must, therefore, inquire how matter could have been expelled from the sun as far as the orbit of Pluto. The velocity required at the sun's surface would have to be nearly the parabolic one, *i.e.*, such as would make the matter leave the sun's influence altogether, something like 500 km. per second. The sun's actual activity shows no signs of such velocities, nor does that of any known star except perhaps novae—also a catastrophic and temporary phenomenon. We must formulate a theory, therefore, to provide for external forces capable of producing such velocities as this near the sun's surface. Buffon's theory provides for such forces; it is hard to see how they could be achieved except through the approach of a passing star to the sun, in such a way as to share its momentum with the outer layers of the sun.

5.2. The theory breaks up into two forms, according as it is supposed that there was merely a close approach or an actual material collision. The former alternative is known as the tidal theory, and was stated first almost simultaneously by Jeans[13,14] and Chamberlin.[15] The latter is in effect that of Buffon but was revived again by Bickerton in 1880 and by myself in 1929. According to the tidal theory the disruption was due to gravitational forces. If a star larger than the sun came sufficiently near, it would raise large tides in the sun; Jeans[16] shows that, if they became large enough, the sun would become unstable and would break up. The complete solution is mathematically prohibitive and we have to proceed by considering extreme cases. In the first, the star is supposed to have approached the sun

and receded so slowly that at any stage the sun had time to take up approximately the equilibrium form under the tidal forces; this is called by Jeans a *slow encounter*. In the other, the process is so rapid that the whole action of the star may be treated as an impulse. This is called a *transitory encounter*. The actual encounter would be *intermediate*. Jeans finds that in a slow encounter there are two possibilities, according to the degree of the sun's central condensation. If it is taken to be nearly homogeneous, it is distorted into a prolate spheroid pointing toward the star, which becomes unstable when the elongation becomes great enough, and breaks up into a number of masses of comparable diameters. This would correspond to the formation of double stars but not to that of planets much smaller than the primary. If it is strongly condensed toward the center, on the other hand, it develops a conical point on its surface, out of which matter flows toward the star and produces a narrow filament. Jeans considers the system of Emden models, which give as one extreme, with γ (the ratio of the specific heats) $= 1.2$, the case of complete concentration to the center; and, with $\gamma = \infty$, the homogeneous case, and estimates (by analogy with the problem of breakup through rotation, which he works out more fully) that the transition comes when the ratio is about 2. For a star in radiative equilibrium the distribution of density corresponds to $\gamma = \frac{4}{3}$, so that the sun would break up by emission from a point.

5.3. In an extreme transitory encounter the sun is in any case elongated in the direction of the line of centers of the two bodies when they are nearest, and rupture occurs if the velocity generated is great enough. But then the ejected matter would simply fall back into the sun when the star had departed, and no planets would be formed. It appears that the actual encounter could not satisfy either condition well. But the ejected matter in a slow or intermediate encounter would be deflected sideways by the attraction of the star on it after it had left the sun and could acquire enough angular momentum to miss the sun on the return journey. The relative path of the two main bodies would be strongly curved; when the star had receded to twice its minimum distance, the direction of motion would already be turned through 45° if the orbit were parabolic. The last condition would be approximately satisfied, since the velocities required are in any case much greater than ordinary stellar ones and must be attributed to the mutual attraction of the two bodies. If the relative velocity at minimum distance is many times that at a great distance, it can be shown that the orbit would be approximately parabolic.

On the tidal theory, only the lighter body can be broken up; to suppose that both are leads to inconsistencies. None of the ejected matter could be carried off permanently by the star, the relative velocity being too great, but probably a great deal would be lost to both bodies.

5.4. The next stage is to see how the ejected matter could condense into planets. If it had no mutual gravitation, it would merely extend longitudinally on account of the differences of velocity between its parts, and to this extent we may regard the formation of a long narrow filament as the immediate result. The issue is complicated, however, by mutual gravitation. Such a filament would be longitudinally unstable; any small excess of mass in part of it would mean a local excess of gravity, which would make more matter collect there. It seems that this instability would develop immediately on emergence from the sun and, strictly, the filament would never exist as a single body, the ejected matter collecting into lumps as fast as it emerged. Indeed, if this were not so, its gravitation would never be strong enough to control the velocities of extension already acquired. The distance between consecutive condensations can be estimated and indicates masses in reasonable agreement with those of the planets.[12]

5.5. The satellites are explained by Jeans as due to further disruptions of the planets by the solar tides at their first perihelion passage. This theory, however, meets with a difficulty in the case of the moon. It has already appeared that, if the disrupted body were strongly condensed to the center, the ejected bodies would be numerous and of small mass in relation to the primary; if it were fairly homogeneous the masses would be few and comparable in size. Jeans, therefore, suggests that the satellites of the larger planets, which are small compared with the primaries, were formed while the primaries were still largely gaseous, but that the earth was broken up when it was already largely liquid. But in these circumstances the earth's density would be so high that it could not be broken up unless it approached the sun within about a quarter of the radius from the latter's surface. The same applies to Mars. As this planet is still smaller than the earth, we should expect it to have been more fully condensed at its first perihelion passage and therefore that its satellites, if any, would have diameters in the neighborhood of 1,000 km., instead of about 10 km. as they actually are. Neither difficulty is fatal, however, and other origins can be suggested for both the moon and the satellites of Mars, without abandoning the theory for those of the great planets.

5.6. The rotations of the sun and the planets are explained on the tidal theory as due to the infall of some of the ejected matter on its

first approach to the primary after acquiring a certain amount of angular momentum. It is here that the tidal theory, after its remarkable successes up to this point, meets its great difficulty. The angular momentum of a particle of mass m is $h = m(fMl)^{\frac{1}{2}}$, where f is the constant of gravitation, M the mass of the primary and l the semilatus rectum of the orbit. For an elongated orbit, l is nearly twice the minimum distance, and for reabsorption the minimum distance must be less than a, the radius of the primary. Hence for a reabsorbed particle $h < m(2fMa)^{\frac{1}{2}}$.

As we know the present rate of rotation, we know the total angular momentum required, and this equation gives a minimum value for m, the mass reabsorbed. For the sun it proves to be about the mass of Jupiter. But for Jupiter it is 0.07 times the mass of the primary and 400 times that of the surviving satellites. Similar considerations apply to the other great planets and to Mars. The earth, Venus and Mercury need not be considered on account of the probable influence of tidal friction on their rotations since they were formed. Now it seems beyond any reasonable expectation to suppose that, of all the matter ejected by Jupiter, only 1 part in 400 escaped reabsorption. For our most favorable estimate the reabsorbed matter has been supposed to have had a pericentral distance equal to the radius of the planet. Yet the amount with a pericentral distance greater than the radius of the planet is only $\frac{1}{400}$ of the whole; insignificant room is left for the departures from the mean value that must have existed on any tidal theory. If we adopt a smaller mean pericentral distance for the returning matter, the angular momentum per unit mass is also diminished and the fraction supposed to have missed the primary on return is made even smaller.

Distension of the great planets at this time makes a slight improvement. But the angular momentum required is the present angular momentum; that capable of being carried by a given mass on returning is only proportional to the square root of the radius. To decrease the reabsorbed mass to that of the surviving satellites, we should have to suppose the primitive radius of Jupiter to be 1.6×10^5 times as great as at present—an utterly impossible distension.[17]

The tidal explanation of the rotations therefore meets serious difficulties, and a more potent cause of rotation seems called for.

6. Such a cause is provided if in our theory we replace the mere close approach of the sun and star by an actual collision. The need to provide a special explanation of rotation really arises from a fundamental theorem of classical hydrodynamics, to the effect that the circulation in a given circuit of particles within a fluid cannot change with the time. In other words, if vorticity is absent to begin with, it

cannot be generated by any forces, such as tidal forces, that are derivable from a potential. This remains true in a viscous fluid, so long as there are no solid boundaries to introduce discontinuities of velocity. If there were no cause of new rotation, the sun and the major planets would have simply kept the rate of rotation of the primitive sun, apart from small changes due to change of density; we cannot explain why Jupiter rotates 50 times as fast as the sun does. The real or viscous fluid can acquire vorticity in two ways that are not contemplated in the classical theorem. One is that, with independent sources of heat, the density may not be a function of the pressure alone; on this principle the science of meteorology depends. But there seems to be no reason to suppose that it is relevant here. The other is that vorticity may be generated at a boundary and diffused into the interior. The usual practical case is where the fluid flows over a solid boundary; but if two fluid masses collide there is a discontinuity of velocity across the surface of separation, corresponding to a local infinite vorticity, which can then be diffused through a layer on both sides of the surface of separation. The discontinuous change of velocity across the interface immediately becomes a continuous transition, which becomes spread through a thicker and thicker layer the longer the bodies remain in contact. Such a transition is a layer of intense vorticity, which is what we need to account for the generation of rotation.

The nature of the flow near an interface may be of two types, laminar and turbulent. In the former the flow at all points of the layer is nearly parallel to the boundary and in the direction of the main flow. This occurs in capillary flow. In turbulent flow the mean motion still satisfies these conditions, but there is superposed on it an irregular movement in all directions. The latter state is the commoner. It occurs if the product of the difference of velocity across the region of transition into the thickness of the region exceeds about 1,000 times the kinematic viscosity of the material; and with ordinary viscosities this is satisfied for quite moderate velocities and linear scales. There can be no doubt that in the conditions of impact at stellar velocities it would be satisfied. This is fortunate, for in turbulent flow the frictional stress across an interface is independent of the viscosity, which is the quantity we know least about. It is equal to $0.002\rho v^2$, where ρ is the density and v the difference of velocity across the interface, in this case the relative velocity of the sun and star. The 0.002 is a numerical coefficient determined by experiment.* We

* Really only for flow over solid boundaries on a meteorological or tidal scale or less. Recent advances in hydrodynamics may lead to results about whether it will still hold on the scale of the present problem. Meanwhile, we can use it as the most natural hypothesis.

can estimate the thickness of the layer of transition easily. Suppose that the collision was an ordinary one, *i.e.*, not a head-on collision or a mere glancing contact. Then the star would slide over the sun through a distance of order a, the radius of the sun, and the time taken is of order $t = a/v$, which is about half an hour. The tangential momentum communicated per unit area is therefore of order $0.002\rho v^2 t$ or $0.002\rho va$. Now this corresponds to the change of momentum of the material in the layer of transition, and the average value of this will be about ρv per unit volume. Hence the thickness of the layer needed to carry the momentum will be of order $0.002a$, and the mass of the matter in the layer will be of the order of $\frac{1}{500}$ of the mass of the sun. Now, as the star receded, this matter would trail after it and would cease to be part of the sun. Thus we should expect the ejected matter to have a mass about $\frac{1}{500}$ of that of the sun. This is in good agreement with the total mass of the planets.

We can also estimate the total angular momentum communicated to the sun during the encounter and that of the ejected matter about its center of mass. The former turns out to be enough to make the sun rotate in a time of the order of 30 days; the actual time is 25 days. The latter implies that, if the whole of the ejected matter were collected into one sphere of the density of the sun, it would rotate in about 8 hr.; Jupiter and Saturn rotate in about 10 hr. At all points the agreement is better than might have been expected, seeing that none of the estimates claims to be more than an order of magnitude. We have seen that a theory of the tidal or collision type seemed to be required to explain the existence of small bodies in the system and the distances of the more remote planets. The tidal theory, when developed constructively, proved capable of accounting directly for several other features, but met with difficulties over the rotations of the planets; and these seem to require material collision to explain them. Now, when a theory of collision is developed, it is found, by simply taking the collision to be as ordinary as possible, we obtain three quantitative predictions about the state of the system, which all agree with the facts, and none of which is even hinted at by any other theory. It seems, therefore, that the theory demands very serious consideration. On the other hand, the agreements are in order of magnitude and not the result of an accurate formal investigation. However, they seem to be as much as can be expected with present data, and closer investigation, which would certainly be very difficult, would evaluate only the mass and size of the star and the extent of the collision, which are not the things that we most urgently need to know.

6.1. The matter dragged out from the sun would be originally in the form of a ribbon, which would tend to oscillate about the circular form. Longitudinal instability would set in as fast as the matter emerged, as on the tidal theory; thus detached masses would be formed. The remaining problems concern the mode of condensation into planets, the formation of satellites and the reduction of the early eccentricities of the orbits.

6.2. The ejected matter, having previously been compressed by the sun's gravitation and being released so as to be subject only to its own, would begin by expanding adiabatically. It appears that if its ratio of specific heats was under $\frac{4}{3}$ it could never stop expanding so long as it remained gaseous. A monatomic or diatomic gas, however, would reach a maximum distension; the radius attainable by Jupiter would be about ten times that of the sun. But with actual materials, liquefaction would set in first on account of the adiabatic cooling; the radius attained is very sensitive to the ratio of the specific heats, but for a monatomic gas would be about 4 million kilometers for Jupiter. At this point, further investigation is desirable because allowance has not been made for the effect of pressure on the boiling point. It is possible that in the central regions condensation might begin much earlier. In any case, however, condensation to the liquid and solid states must have played an important part and would apparently have begun in a week or so from release. Further cooling would take place by radiation from the surface; again the estimates available are subject to modification, but it appears that the original thermal and gravitational energy could be disposed of in a time of the order of months or years. While this was going on, the drops formed by liquefaction would settle to the center and form the planets as we know them. The more volatile constituents, however, might fail to liquefy and ultimately be lost to the planets. The lighter the planet the greater the loss; this seems to explain why the larger planets have the lower densities. Condensation would apparently be well advanced in all the planets at their first perihelion passage.[18]

6.3. The formation of satellites remains a difficulty. Tidal disruption of the primaries at their first perihelion passage is possible, but as an explanation of the satellites it meets with a dilemma. If the planets were already highly condensed, tidal disruption would give satellites whose diameters were moderate fractions of those of their primaries; the only such satellite is the moon and, as we have said already, the approach of the earth to the sun would have to be dangerously close. On the other hand, if the satellites were still distended, it seems unlikely that any of them would have been able to hold themselves

together in the gaseous state and certain that the majority of them could not. It seems most probable that smaller condensations arose in the matter as it was originally ejected and that the original expansions of the planets extended beyond the present orbits of most of the satellites; in such a case the satellites, even if gaseous, would be surrounded by gas at a finite pressure, and their expansion would at any rate be delayed. But then there is a difficulty about their rates of revolution. If a satellite were formed in this way, it would presumably have the mean velocity of its surroundings; we require it to have the velocity corresponding to motion in a circular orbit about its primary. To satisfy both these conditions requires that the rotation should have been that of a Laplacian nebula on the verge of breaking up and requires much more angular momentum than the subsystems have at present, unless condensation to the center were already nearly complete. On the other hand, we cannot ignore the type of condensation that produces snowflakes. This does not depend on gravitational forces but simply on the actual gas pressure exceeding the vapor pressure of the material at the actual temperature. It might, so far as can be seen at present, happen even when most of the planetary matter had collected to the center. We lack relevant data at present about the low vapor pressures involved. There is, however, at least one satellite, Mimas, that suggests such an origin. Hepburn[19] shows, from the mass and luminosity of this body, that its density cannot exceed about 0.3, which is less than that of any known homogeneous solid except hydrogen and helium, neither of which is likely to be solid at the temperature of the satellite. This seems to indicate that Mimas is not a uniform solid, such as would be formed by condensation through the liquid state, but a loose aggregate of crystals, such as would be formed by direct condensation from a gas at low pressure.

7. There is an alternative theory of the origin of the moon, due originally to Sir G. H. Darwin and depending on his development of the theory of tidal friction. We have seen that by this theory we can, given enough time, trace the moon back to the time when it revolved in about 5 hr. and the earth rotated in the same period; before that the moon can apparently not have existed as a separate body. The natural suggestion is that, just before this, it was part of the earth, which in some way separated into two bodies. Darwin noticed that if the moon was combined with the earth, the angular momentum being retained, the period of rotation would be about 4 hr. Now this is near a critical speed for two possible events in the history of a fluid mass with the density of the earth. A homogeneous fluid mass with a slow rotation has the form of an oblate spheroid, the

Maclaurin spheroid. The oblateness of this increases with the speed of rotation, and at a certain critical speed another type of form becomes possible, an ellipsoid with three unequal axes, called the Jacobi ellipsoid. For higher speeds the Maclaurin ellipsoid is unstable if the fluid is viscous, but the Jacobi one is stable. At a still higher speed the Jacobi ellipsoid in turn becomes unstable, a furrow forms near one end and deepens rapidly and then the mass breaks up into two. It was thought for a time that the moon might have been formed in this way, but Moulton[20] showed that the rate of rotation could not have been enough to make the Maclaurin form unstable, much less the Jacobi form.

Darwin, however, noticed also that 2 hr. is near the period of a free oscillation of a fluid mass with the density of the earth, such that two diameters at right angles lengthen and contract in turn. This is just the type of oscillation that would be produced by the solar tides. Further, the period of the tidal oscillation is half that of rotation, so that the periods of the free and forced oscillations agree. This is the condition for the phenomenon of resonance; the tidal movement, beginning small, would increase steadily in amplitude, and there is no obvious limit. But if it exceeded a certain amount the mass would become unstable and break up. This is the resonance theory of the origin of the moon. If we treat the mass as homogeneous, it fails for the same reason as the rotational theory; for Bryan showed that resonance would become possible only at the same speed as made the Maclaurin form unstable, a speed that Moulton afterward showed could not be attained. But if we allow for the earth's central condensation, we find that resonance is possible at a somewhat lower speed of rotation, which seems attainable (Ref. 1, Chapter 3).

However, allowance for the earth's heterogeneity, though helping resonance in one respect, destroys it in another. The earth's central core is shown by seismology to have a sharp boundary capable of producing definite reflected waves, implying a definite discontinuity of material, usually interpreted as a rocky shell upon a core of liquid iron. In the liquid state, tidal oscillations would imply friction over this boundary, which would limit the amplitude attainable by resonance. It appears that the amplitude would cease to increase when it was only about one-twentieth of the radius, and there is no question of instability.[11]

A further difficulty in the theory was suggested to me privately by E. W. Brown. The ordinary theory of resonance treats the equations of motion as rigorously linear. In most actual problems this is an approximation involving the neglect of higher powers

of the displacement. When these are retained, a great variety of phenomena can arise, but the ordinary theory shows only that the amplitude, with a suitable agreement of periods, will rise to such an amount that the higher terms must be retained. It seems that the effect of these is always to limit the amplitude attainable, and it is probable that even without friction the amplitude of the tides in the earth would never become enough to give instability.

It seems, therefore, that the moon cannot have been formed from the earth since the latter became liquid, except perhaps on the lines suggested by Jeans. In any case the moon is practically the same age as the earth, and its origin was one of the events arising directly from the original catastrophe.

8. As we have indicated, the planets would lose a great deal of their mass after ejection, certainly much of the more volatile constituents and possibly much of what had time to condense into drops. This matter would proceed to revolve about the sun, in the first place in highly eccentric orbits like those of the primitive planets. Thus the sun would be surrounded by a tenuous gas, mixed probably with a number of small solid particles. Both gas and particles would afterward resist the planetary motions; it remains to be seen which will play the more important part. In either case, on account of the differences in the periods, the motions in any small region would quickly become distributed at random about some mean, this mean being a transverse velocity in the same sense as the revolution of the planets.

8.1. The solid particles have been thought to be still represented in the present system by meteors. Our immediate concern, however, is not with the meteors as such but with the extent of the effect of meteoric aggregation on the planets. The result of any meteoric impact on a planet would be that the velocity of the combined body after impact would be a weighted mean of those before impact, and as the average radial velocity of the meteors would be zero the tendency would be for the aggregation to reduce the radial velocities of the planets and therefore the eccentricities of their orbits. The same can be shown to be true of the variations in the transverse velocities. For a substantial reduction to be obtained, the total mass of the meteors picked up by a planet must be several times the original mass of the planet. But at this point the theory proves to be internally inconsistent. If we consider any individual meteor (or planetesimal, in Chamberlin's language), the probability that it will strike another meteor in a given time is to the probability that it will strike a planet

in the ratio of the total surface of the meteors to that of the planets. If, then, the total mass of the meteors was comparable with that of the planets, their surface must have been much greater on account of their finely divided state. The meteors would therefore have collided with one another and volatilized themselves before the planets had had time to grow appreciably. It appears, therefore, that the solid particles, however interesting intrinsically, cannot have influenced the history of the planets, which have now practically the same masses as when they were formed.

The present rate of accumulation of meteoric matter on the earth would cover it to a depth of the order of 2×10^{-5} cm. per 100 million years.[29] But in any case the high eccentricities of meteoric orbits make it very unlikely that present meteors have anything to do with the original solid particles, if these existed.

8.2. The gaseous constituents, on the other hand, are capable of producing the desired effect. It can be shown that in a gaseous medium the mean velocity at any point would be practically that of a planet in a circular orbit at the same distance, and friction would tend to damp out the variations of the planets' velocities. The rate of reduction of the eccentricities would be proportional to the density of the medium. On the other hand the medium would itself degenerate owing to viscosity. The only permanent motion of a gas or liquid is such that every part takes the same time to revolve about an axis; but in the system we are considering the time of revolution was dominated by the sun's attraction and increased with the distance from the sun. Thus internal friction caused a secular change in the motion of the gas. The more distant parts had their motion accelerated by the faster moving inner ones and were in consequence made to go farther and farther off, whereas the nearer ones as a result of the same process were made to approach the sun and were reabsorbed into it. Thus the gas would gradually disappear, the time taken being proportional to its original density. This time must also be the time needed to reduce the eccentricities of the orbits to their present values; for, if the medium lasted longer than this, it would either be still conspicuous or have made the orbits more nearly circular than they are, and if it did not last so long it would have left them more elongated than they are. But the time needed to produce this effect is inversely proportional to the original density. Consequently the relation that the two times shall be equal gives a means of determining the unknown original density of the gas and the time taken. The latter is found to be of the order of 1,000 million years.[1] It appears that the age of the

earth is between 1,500 and 3,000 million years, so that the agreement is as good as can be expected. It is possible that the zodiacal matter is the last remnant of the medium.

8.3. The satellites, again, would be affected by a resisting medium in much the same way as the planets were. But a serious mathematical complication has hitherto made any adequate investigation of the effects impossible. The gravitational attraction of the planets would draw the gas toward them, producing a local increase of pressure and density. There would be a limit to the pressures attained, and thus there would be steady condensations about the planets; these would have to be pushed through the rest of the medium by the planets, the resistance thus being greatly increased, since in all major planets, except Mercury, the diameter of the condensation would much exceed that of the planet.* But the motion of a satellite about a planet, as perturbed by the sun, is one of the most difficult dynamical problems ever solved, and the motion of a continuous medium in similar circumstances is necessarily a still harder problem. Consequently, though we can estimate the order of magnitude of the diameters of the condensations and can disprove various propositions that might be thought likely about the motion of the matter within them, we cannot determine their sizes or their internal motions accurately. The size of the condensation being an important factor in determining the resistance of the medium to the motion of a planet, we cannot apply the method of determining the age of the system from the eccentricity of the orbit to any planet except Mercury. We should expect the sizes of the condensations about the great planets to be such that all their satellites, except the four outer ones of Jupiter and the two outer ones of Saturn, were within them from the start. It is in accordance with this that the inner satellites move in orbits imperceptibly different from circles and in the planes of their pri-

* Brown,[9] p. 465, appears to have misunderstood this point, for he suggests that the resistance implies an accession of mass by the planets. The resistance would be of the same type as that experienced by a rifle bullet traveling through air; the pressure is increased in front and reduced behind, and the resultant thrust opposes the motion. But the individual molecules leave the neighborhood of the bullet as fast as new ones approach it, and there is no accumulation. In the case of a large planet, gravity will superpose on this effect an increase of pressure all round, but this is independent of the time and again requires no accumulation. There would be accumulation only if there were some means of condensing the gas into liquid or solid on the planet's surface; then the pressure of the newly condensed material disappears and is no longer available for holding other material off, so that a steady approach to the planet would be possible. But the medium is here supposed to consist of a permanent gas, mainly hydrogen, and the question of change of state does not arise.

maries' equators, apart from small disturbances due to perturbations. But it is also probable that the medium within the condensations moved so as to offer a small steady resistance to the motions of the satellites and make them approach their primaries; this has not been proved but would explain several curious facts about the distances of the satellites if it is true.

There are a few anomalous satellites, which move about their primaries in the opposite direction to the revolution of the planets; these are the outermost two of Jupiter, the outermost of Saturn, all four of Uranus and the single one of Neptune. In the case of Uranus, but not of Neptune, the rotation of the primary is itself in the anomalous direction, and the satellites move as might be expected if their motion had in some way been determined by the rotation of the primary. On the nebular theory a single rotation or revolution in the wrong direction would present a difficulty almost impossible to explain, but on the tidal and collision theories exceptions are hardly surprising, considering the possibilities of irregularities in the motions after ejection. The retrograde satellites of Jupiter and Saturn, associated with direct ones in the same subsystems, can hardly be explained by the normal mechanism. On the other hand there seems to be no reason why they should not have been at one time independent planets and been captured by their present primaries. The same may apply to the satellites of Mars.

9. It should be remarked, as Nölke has pointed out,[4] that the angular momentum of the medium per unit mass is proportional to the square root of the distance. Thus, given a finite total angular momentum to begin with, there is a limit to the mass that can be driven beyond any assigned distance. The medium would spread slowly outward, but the ultimate fate of practically all of it would be absorption in the sun. It has appeared that the total mass of the medium was probably comparable with that of Jupiter, and the angular momentum carried back to the sun would be that of a body with the mass of Jupiter revolving at the sun's equator. This turns out to be about two-thirds of the sun's angular momentum, so that a considerable fraction of the sun's rotation is accounted for by the return of the medium.[22] Our knowledge is not yet sufficient to say how much of it is due to this and how much to friction during the original collision; the contributions seem to be of the same order of magnitude.

10. The large eccentricities and inclinations of the asteroids present some difficulty, since these small bodies would be expected to be more affected than Mercury by the resisting medium. Their small size and their concentration between the orbits of Jupiter and Mars have

suggested to many that they may be the result of the disruption of a small body by explosion or the tidal action of Jupiter; if so, they may not have existed as separate bodies till a late state in the development of the system. On the other hand, on Brown's theory of the meaning of the failure of the methods of celestial mechanics over long intervals of time, their high eccentricities and inclinations may represent an approach to equipartition of the anomalous motions.

Meteors and comets have not yet been included in any theory of the origin of the system (but see Brown[9]). It seems to me, considering the large surface that the asteroids must have, that a number of them may have undergone collision and shattered one another. A terrestrial origin for meteorites has been suggested, on the ground that a body expelled from the earth would have a greater chance of striking the earth again than one not so expelled; but this argument loses its force when we see that the solar system has existed about ten times as long as the earth would need to sweep up most of its neighborhood. The appearance of stony meteorites suggests that they solidified slowly and are, therefore, parts of a larger body.

11. The foregoing account was originally written in 1933 and represents the position at that time. Since then, two serious criticisms have been made by H. N. Russell and H. P. Robertson. Russell's objection is that the collision theory does not provide enough angular momentum for the planets.[30] It is the opposite objection to the one against the nebular theory, which provides too much. The argument about the rotation of Jupiter can be applied to the original encounter. The star would have to approach to a distance comparable with the sun's radius, and all ejected matter would begin at that radius. The star's velocity would change direction rapidly after the encounter, and it would be only while the distances involved were all comparable with the sun's radius that any important communication of angular momentum about the sun to the ejected matter could take place. Subsequent changes would not affect the angular momentum, and the angular momentum per unit mass would be of order $(2fMa)^{\frac{1}{2}}$, where M is now the mass of the sun and a its radius. That in a circular orbit of radius b is $(fMb)^{\frac{1}{2}}$. It appears therefore that with all the rounding up of the orbits that is possible the average distance of the planets would be of the order of twice the sun's radius. The distance of Jupiter is about 1,000 times the sun's radius. The discrepancy seems beyond any possibility of reconciliation through a consideration of the uncertainties of the estimates. The resisting medium would be concentrated near the sun from the start, and resistance would be important only when the planets were near perihelion. Thus it would

have little effect on the perihelion distances, but the loss of energy at every approach to the sun would reduce the distance at the next aphelion, and when the orbits were finally rounded the mean distance would be comparable with the original perihelion distance, not the original mean distance.

Russell offered a suggestion that at the time of the encounter the sun was a binary star, the companion having a distance comparable with those of the great planets, and that the wandering star collided, not with the sun, but with the companion. In this case the requisite angular momentum to account for the planets' distances is already available, and since, as he remarks, about one-tenth of the stars are known to be binaries such a collision is almost as likely as one with the sun. He discarded this hypothesis, however, for various reasons, the chief of which was the difficulty of disposing of the companion after the collision. The question was, however, reopened by R. A. Lyttleton.[31] He showed, by remarkably neat methods, that in such an encounter the exchange of momentum between the companion and the wanderer could leave both with velocities that would make them escape from the sun and yet that much of the ejected matter connecting them would have velocities below the velocity of escape and would therefore stay with the sun. The solar system now has three parents instead of two. This hypothesis has all the advantages of the collision theory but avoids its chief defect, and it requires very serious consideration. Lyttleton has also obtained a remarkable result about the recently discovered outer planet Pluto, and Triton, the anomalous satellite of Neptune.[32] These two bodies are of comparable size, and Pluto at its perihelion is within the orbit of Neptune, so that close encounters between Neptune and Pluto may be expected. Lyttleton shows that in suitable circumstances Pluto could encounter Triton, in such a way that both would be converted into direct satellites of Neptune. Now the process is reversible, and the argument also shows that if Pluto and Triton were once both direct satellites of Neptune, a close encounter between them could have led to the reversal of the orbital motion of Triton and the ejection of Pluto to become an independent planet. The explanation suggested of the anomalous revolution of Triton would remove one of the difficulties of all theories of the origin of the solar system, and there is considerable reason to believe that the event in question has actually occurred. It remains to be seen, however, whether other retrograde satellites can be explained in this way.

Robertson's work[33] concerns the effect of radiation pressure on small bodies revolving about the sun. If in circular orbits, these are crossing the rays of light and heat from the sun and therefore absorb

energy, which will appear from the bodies themselves to be coming from a direction not quite radial. The energy radiated away, however, will travel in all directions relative to the body, and the result will be a radiation pressure mainly away from the sun but with a small component against the orbital motion, which will reduce the angular momentum steadily. In a detailed analysis Robertson shows that a particle 1 cm. in radius and of density 5.5, starting at the distance of the earth, will be swept into the sun in 4×10^7 years; one of radius 0.001 cm. would survive only 4×10^4 years. The life is proportional to the square of the initial distance, so that apparently a particle of radius 1 cm., starting at the distance of Jupiter, could survive for 10^9 years. Nevertheless, Robertson's work makes it doubtful whether a gaseous resisting medium could exist for long enough to produce the reduction of the eccentricities of the planets' orbits that has been required of it. In such a medium, radiation would perhaps be absorbed and re-emitted several times before leaving the system, and the net effect may be equivalent to an increase of the viscosity, which may be calculable. As the total mass of the radiation lost in the lifetime of the earth is of the order of one-quarter of that of the planets, and as the medium, if it lost angular momentum, may have had some of it restored by the planets, the present writer does not think that the medium need be immediately abandoned; but the matter certainly needs examination.

A further difficulty concerns the densities of the planets. It is natural to suppose that the larger nuclei after the original catastrophe were able to retain more light materials than the smaller ones, and this provides an explanation of the fact that the great planets have much lower densities than the terrestrial ones. But, when we compare the inner planets among themselves, we find that in the order Earth, Venus, Mars, Mercury, the moon is the order of decreasing mass and also of decreasing density; the relation between the larger and smaller planets is just reversed.[34] A detailed consideration of the probable distribution of density in them shows that the variation is due mainly to the relative size of the dense core, which occupies half the radii of the Earth and Venus, possibly 0.4 of that of Mars, and is probably absent from Mercury and the moon.[35] To explain this variation would apparently require that the smaller bodies lost iron in comparison with silicates, which would be very difficult to understand. It could be explained if they once formed only one or two bodies and the iron had time to settle to the center before these were broken up; but this would apparently require an evolutionary origin and not a catastrophic one, such as seems to be

demanded by the existence of the asteroids and satellites. The present writer's opinion is that the outstanding problem of the history of our system is to reconcile these two lines of evidence, one pointing definitely to a catastrophic origin and the other to a gradual one.

Further work by Lyttleton[37,38] has paid special attention to this problem. It had been shown by Cartan, in a paper that had been overlooked, that when the Jacobi ellipsoid becomes secularly unstable it also becomes ordinarily unstable. Following out the consequences of this, Lyttleton finds that a body breaking up through excessive rotation would not give a planet and satellite, or a double star, as had been supposed. It would give two independent bodies. This opens a possibility that the planets originally formed only one or two bodies, which broke up through rotation, and this has the merit of being the first suggestion that appears to account for the different densities of the terrestrial planets.

12. Tidal friction has already been mentioned briefly; in the case of the earth it can be evaluated from present data.[1,23] But tidal friction on the earth occurs mostly in shallow seas around the continental margins, and it appears that bodily tidal friction is small in comparison.[24] No other planet, except perhaps Mars, has both satellites and shallow seas, but Mercury, Venus, the moon and several satellites show signs of the effects of tidal friction, which must therefore have been bodily. The tides raised in the planets by the sun tend to make them rotate in the period of revolution, thus always keeping the same face to the sun; Mercury does so, and the rotation of Venus, though not known with accuracy, is slow. Other planets have not had time to be much affected by the smaller solar tides in them. The moon and some other satellites also keep the same faces to their primaries. Some satellites, notably Phobos and JI, would be rapidly affected by tidal friction if their primaries showed any imperfection of elasticity. On the supposition of bodily tidal friction, Darwin followed out in great detail the evolution of the orbit of the moon and the inclination of the earth's axis to the ecliptic; it would be desirable to know to what extent his results need modification in our present state of knowledge.

The moon shows a curious excess ellipticity. The difference among its moments of inertia is about sixteen times what it would be had it solidified at its present distance and with its present speed of rotation; as if it had solidified, or at any rate last adjusted itself to a state of hydrostatic pressure, when about one-third as far from the earth as it is now.[1,25,26,36] The extra load corresponding to this excess ellipticity must have been supported by the strength of the moon's

materials for a time comparable with the age of the earth and provides a piece of positive evidence against the common opinion that rocks will undergo permanent deformation under any stress, however small, provided that it is maintained long enough. There is purely geophysical evidence that the earth's rocky shell at great depths can stand stress differences of the order of 5×10^7 dynes per square centimeter, but it is sometimes suggested that these are not permanent. A strength of the same order is indicated for the moon, and there is no room for doubt that it is permanent.

Geophysics provides cosmogony with several of its fundamental data; cosmogony repays it with one. The earth was once liquid. In this state its free iron and materials soluble in it settled to the center, leaving the silicate shell outside. Since then the shell has become solid, but the core remains liquid to this day.

SUMMARY

It appears from various lines of investigation that the solar system some thousands of millions of years ago must have departed widely from its present state and that the earth cannot have existed as a separate body for more than about 3,000 million years at the outside. The sun at that time must have been in nearly its present state. The smaller bodies in the system cannot have been formed by slow condensation from the gaseous state, and the present extension of the system implies a more violent disturbance than any we can suspect from the present state of the sun. It appears that the breakup of the sun through the tidal action of a passing star can account for many of the features of the system but fails to account for the rotations of the planets. If, however, the star actually collided with the sun, a cause of rotation is provided, and such a theory gives estimates of the total mass of the planets and the rates of rotation of the sun and planets which agree with the facts. There is a further difficulty in the collision theory in accounting for the production of enough angular momentum, but this can apparently be avoided on the assumption that the sun was originally a double star and the encounter was not with the sun, but with the companion. The later development of the system is sketched, and points where further investigation is required are indicated.

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CHAPTER III

RELEVANT FACTS AND INFERENCES FROM FIELD GEOLOGY

REGINALD A. DALY
(Professor of Geology, Harvard University)

INTRODUCTION

The face of the earth records ceaseless molding by solar and lunar influences during more than 1,000 million years, but the major lineaments have been fashioned by internal forces. If we can read the face aright, we shall be able to learn something substantial about the nature of the body. In fact the world map, topographical and geological, is the primary base of supplies in the mental campaign against the mystery of the deep interior. The observations of the field geologist are not only suggestive; they are also controlling to a degree, and ideas regarding the invisible layers of the globe must be consonant with the facts of the surface. For example, the face of the earth is rugged; no theory of the interior is good that fails to show how this topography is supported. Again, temperatures computed for the depths should agree with the thermal gradient at the surface. And a third illustration of the principle is found in the fact that the natural choice of materials ruling at depth is fixed by the range of types among accessible rocks, supplemented, it may be, by the range of types in meteorite and moon. Thus the geophysicist has to become a geologist, just as a geologist, to be worthy of the name, does his best to master methods and results of geophysical study. Furthermore, since the present state of the earth's interior is the evolutionary product of its states in the long past, the geophysicist asks the geologist to furnish all possible information that can be derived from paleogeography in the broadest sense of the term.

This chapter enumerates the principal facts determined by geologists working in field or laboratory and draws therefrom some direct inferences concerning the inaccessible depths. The matters under discussion may be listed under six heads: (1) the visible topography, composition and structure of the lithosphere; (2) certain clues derived from analysis of the deformations of the rocky units; (3) the validity of speculations involving (*a*) recent, pronounced instability of the

intertropical belt, (*b*) recent, pronounced instability of continental shelves, (*c*) extensive migration of the earth's poles of rotation, and (*d*) triaxiality for the earth's figure; (4) specially important facts from the field and laboratory study of the igneous rocks; (5) a summary statement of the time during which our planet's interior has come to its present condition and during which potential energies have there accumulated or been converted into kinetic energy; (6) a plea for ever more systematic testing of mental models of the earth by those trained in physical geology.

GENERAL TOPOGRAPHY OF THE GLOBE

The continent of physical geology includes its submerged shelf and outer slope and is therefore bigger than the continent of the school

TABLE 1

DIMENSIONS OF EARTH, LANDS AND SEAS

	Area, sq. km.	Average depth, km.	Average height, km.	Volume, cu. km.
Whole earth	510,100,000	.....	....	1,083,000,000,000
Whole ocean	361,160,000	4,117	....	1,486,900,000
All land	148,940,000			
Pacific Ocean	165,200,000	4,282	....	707,500,000
Atlantic Ocean	82,400,000	3,926	....	323,600,000
Indian Ocean	73,500,000	3,963	....	291,000,000
Arctic Ocean	14,090,000	1,205	....	17,000,000
Seas	25,970,000			
Asia	44,134,000	.....	960	
Europe	10,009,000	.....	340	
Africa	29,834,000	.....	750	
North America	24,063,000	.....	720	
South America	17,788,000	.....	590	
Australia	8,901,000	.....	340	
Antarctica	14,169,000	.....	2,200	

atlas. The whole relief may be regarded as an enormous curved buckler, resting on the body of the earth and flanked by the deep ocean. There are five such bucklers: Eurasia-Africa, Australia, Antarctica, North America and South America. Each has its shallow, overlapping seas of transgression, like the broad Sunda Sea between Sumatra and Borneo, the Sahul Sea between New Guinea and Australia, Bass Strait between Tasmania and Australia, Bering Sea, Hudson Bay and the Baltic Sea. The continental lands are also interrupted by the much deeper seas of ingression, like the Sea of

Japan, some of the great basins of the East Indies, the Red Sea, the Mediterranean, Ross Sea and the Gulf of Mexico.

The ocean, including the satellitic seas of transgression and ingression, covers seven-tenths of the earth and, beyond the outer limits of the bucklers (outside the foot of the continental slope), covers six-tenths. The average depths of water are given in Table 1, which also states the average heights of the continents above sea level.[1]

The existing bucklers are fragments. The Old and New Worlds bear ample traces of wide Paleozoic seas of transgression, but throughout most or all of that era Eurasia, Australia and the two Americas were so thoroughly connected by dry land that it is fair to say that all four formed one buckler, which may also have included Antarctica. At that time, the term *land hemisphere* was even more clearly justified than it is now, and Alfred Wegener's name *Pangaea* for the inclusive Paleozoic buckler seems essentially justified, though the run of its coast line may have differed much from that drafted by Wegener.[2] The conditions for the breakup of Pangaea were prepared far below the surface of the planet, and hence the revolutionary event is one of the supreme facts to be considered by the geophysicist.

Although part, probably the greater part, of each of the Atlantic and Indian ocean basins originated in post-Paleozoic time, the geological evidence favors the view that the Pacific has represented the *water hemisphere* ever since Pangaea was developed, *i.e.*, since a date well back in the Pre-Cambrian era. In general, geologists are skeptical of the biologist's *ad hoc* hypothesis that a vast continent has foundered in the area of the central Pacific.

COMPOSITION AND EXTENT OF THE VISIBLE TERRAINS

The bucklers.—Down to a limited depth each continental buckler is essentially a *Basement Complex* covered with veneers of younger rocks and locally invaded by intrusions of post-Archean dates. Of course the word *Basement* is here used in a relative sense and refers only to the visible or clearly inferable part of the foundation on which the veneers rest. To this lower course of the colossal masonry the field geologist sees no bottom. The word *Complex* expresses his honest confession that he is far from understanding the structural turmoil in the principal terrain of the bucklers.

Except for a few patches of veneer, Finland is surfaced by the Basement Complex. In that extensive land J. J. Sederholm measured the areas covered by the different kinds of rocks and then computed the average composition of the terrain at the surface.[3] With due allowance for the comparative thinness of the volcanic masses, chiefly basic,

this average was found to be close to that of granite, with about 70 per cent of silica. The average density of the unweathered rock at the surface is almost exactly 2.70. Reconnaissance and considerable detailed mapping show that in these two respects Finland is a good example of the Basement Complex in every continent.

The areally important veneers are of three kinds: sedimentary, volcanic and glacial.

The sedimentary veneers comprise both flat-lying strata and also less extensive, strongly deformed strata, *viz.*, those folded, thrusted and faulted. For the most part the strata have been derived from the disintegration of the granite of the Basement Complex, followed by deposition of the solid detritus on the floors of the wide seas of transgression and in basins, including the gulfs of the elongated geosynclines.

The volcanic veneers of great extent are made almost wholly of plateau basalt, the maximum single area and maximum observed thickness being respectively about 500,000 sq. km. and about 2,000 m.

The Antarctic icecap, with maximum thickness probably exceeding 2,000 m., covers 13,500,000 sq. km. The Greenland icecap, with the same order of maximum, has an area of nearly 2,000,000 sq. km.

Basement Complex and rocky veneers alike have been invaded by younger magmatic, eruptive masses, granite dominating among the intrusives and basalt dominating among the extrusives.

Oceans.—Most of the larger oceanic islands show, both by their geographical relations and by their composition, that they are, or once were, parts of the continental bucklers. New Zealand, though 1,500 km. from Australia, is no exception, for the two were once joined in the ancient Gondwanaland. The sea bottom between the two is rough, as if underlaid by continental rock of greatly varying thickness. Kerguelen Island and the broad plateau around it probably represent an outlier of Antarctica, 1,600 km. away. Nearly midway between Europe and North America, the Azores have composition and submarine contours suggesting a drowned continuation of the Alpide mountain system of the continent.

On the other hand, the 3,000 islands of the deep sea, representing the only visible rock in half the earth, are all small and apparently all of volcanic origin. Without exception the overwhelming mass of each pile of lava is basalt, which, however, is here and there associated with small outflows of its own derivatives, typified by trachyte. Quartz-bearing and other typically continental rocks are conspicuously absent in all but two or three of these truly thalassic islands. Exceptions to this rule include Easter Island of the open Pacific and Ascension Island in mid-Atlantic.

On Easter Island, M. C. Bandy has recently mapped short flows of rhyolitic obsidian, emanating from old basaltic craters.[4] It is significant that this island, essentially a composite basaltic cone, rests on the Albatross Plateau, whose surface is about 3,500 to 4,000 m. below sea level and thus 1,000 to 1,500 m. higher than the general floor of the Pacific. Both the hypsometry and the existence of what is, chemically speaking, granitic material in the island suggest the possibility that the plateau is an extensive but relatively thin slice of rock of continental type, covering denser rock like that composing the "crust" under the deeper part of the Pacific.

Ascension Island is the top of another huge volcanic complex, almost completely basaltic in composition. While mapping it, the present writer was able to corroborate Darwin's discovery of granitic bombs, thrown up from below the original sea bottom and incorporated in the tuffs and breccias.[5] Like the Azores, Ascension is, as it were, perched on the back of the Mid-Atlantic Swell, a topographic analogue of the Albatross Plateau. Thus there is some evidence that a thin, broad, irregular strip of "continental" rock floors the axial region of the Atlantic through a total length of 11,000 km., or 7,000 miles.

From the soundings of the John Murray Expedition it appears that the floor of the western half of the Indian Ocean is similarly varied by "swells" and narrower reliefs.[6] The Arabian Sea is traversed by the submerged Murray Ridge, which has been mapped as the northern part of the much greater Mid-Indian Ridge. To the westward of this lie the Carlsberg Ridge and the Seychelles Bank. Since the ridge and bank have similar hypsometric relations and since the Seychelles Bank emerges at a few points in the form of granitic islands, we may assume that all these elements in the accidented topography of the Indian Ocean floor are drowned remnants of the Paleozoic Gondwanaland. This assumption may help to explain Vening Meinesz's gravity profile between Aden and Ceylon.[7]

INFERENCES

The world map warrants a few deductions, which have not the sanction of all geologists and yet appear to have much of the authority of observed fact and to offer important clues to the nature of the earth's interior.

1. All known terrestrial rocks originated ultimately from magmas or melts; the geologist has found no direct evidence that the planet contains any definite layer of infallen "planetesimals."

2. Although shallow seas have transgressed and regressed on the bucklers, these have in general stood high above the Pacific floor ever

since an early Pre-Cambrian date. The simplest supposition is that the sial (silica-aluminum-rich material) of Pangaea, intact or fragmented, has kept its height because of the balancing pressure of the denser sima (silica-magnesium-rich material) under the deep ocean. In other words, the principle of isostasy (with regional compensation) seems to have ruled ever since dry, continental land appeared, though, of course, the level of compensation may have changed through a considerable range.

3. The persistence of the buckler relation proves that sialic rock has withstood shearing stress for 1,000 million years; it has considerable ultimate strength.

4. The visible, crystallized sima has strength of the same order, as shown by the prolonged existence of the high-standing basaltic volcanoes and by the sturdy resistance of cliffed basalt and gabbro to the call of gravity for millions of years. Great weakness for the sima at the range of temperatures between the earth's surface and the depth of a few tens of kilometers is out of the question. The bearing of this truth on the theory of continental migration is obvious.

5. The fact that basalt, itself simatic, has been erupted in great volume on buckler and ocean floor alike finds an immediate explanation if this magma is assumed to have risen from an eruptible world-circling layer of the same composition.

6. The nearly perfect restriction of the sial to one hemisphere, a restriction already made in early Pre-Cambrian time, is particularly intriguing to the student of the earth's interior. The old view, that the sial (density about 2.70) is so largely missing in the Pacific hemisphere because it was torn away to form part of the moon does not easily account for the moon's mean density (3.3).

DEFORMATION OF THE LITHOSPHERE

Beneath sea and land is crystalline rock, and crystalline rock at the low temperatures is strong. It is abundantly clear that below this world-circling material there is a much weaker, world-circling layer. These two shells bear the useful names *lithosphere* (rocky shell) and *asthenosphere* (weak shell), respectively. Among the outstanding questions of this book are: (1) the thickness of the lithosphere, and (2) the cause of the weakness in the asthenosphere. Both questions seem to find answers only in a single condition, *viz.*, sufficient increase of temperature with depth. Whether this increase means that the lithosphere is a true crust and the asthenosphere is a shell too hot to crystallize is an old problem, still so speculative that, in presenting the

objective facts, it is safer to refer to the superficial shell as "lithosphere" rather than "crust."

Important evidences for the existence of the two contrasted shells are derived from observed deformations of visible terrains. These changes are of four kinds: (1) broad warpings of the rocky surface of the globe, with no obvious relation to sedimentary or other veneers; (2) basining, genetically connected with the veneers; (3) mountain making (orogeny); and (4) deformations in the Basement Complex.

1. *Ocean deeps and other open geosynclines.*—The mountain arcs bordering the Pacific basin were evidently developed under intense horizontal compression. Many of the arcs, like the Japanese, Philippine, Tongan and Andean, are fronted by ocean deeps, the grandest of the existing "open" geosynclines. There is no reason to doubt that these elongated, water-filled downwarps of the rocky surface were formed by the same horizontally directed pressures as those responsible for fold and thrust in the adjacent chains of mountains and for the arcuate advance of these toward the Pacific. Owing to the comparative recency of the thrusting and downwarping and also to distance from land, these geosynclines have escaped filling with detritus; they are still open and in general show strongly negative gravity anomalies.

It is unlikely that these great downwarps were caused by pure flexure of an initially dead-level lithosphere, as in the instructive experiments of Kuenen[8] and MacCarthy.[9] As pointed out by Smoluchowski and Vening Meinesz,[7] the lithosphere is too weak to carry the required horizontal pressure. Nevertheless, the downwarping under horizontal pressure is inevitable, if the lithosphere is first broken by a throughgoing thrust-fault. In this case, the downthrust part of the lithosphere has to be pushed down by the overthrust part, with downbending farther out to sea. Thus these ocean deeps, like the Ganges trough of northern India, seem best explained as "fault warps."

It may be added that the opposed view, whereby geosynclines are supposed to be due to horizontally directed tension in the earth's superficial layer, is supported neither by the facts of the field nor by laboratory experiments.

The ocean deeps are long but at most only a few hundreds of kilometers in width. Hence they seem to prove relative thinness and real flexibility for the lithosphere. Moreover, since the depression of the solid-elastic lithosphere demanded lateral, horizontal flow of the underlying material, the reality of the asthenosphere is suggested.

2. *Deformations genetically connected with the veneers.*—The suggestion just mentioned agrees with the facts known about geosyn-

clinal prisms, *i.e.*, the sedimentary fillings of ancient geosynclines. The floors of some of the prisms were forced down to a far greater depth than that reached by the rocky surface of the deepest known open geosyncline. This excess of basining is to be largely accounted for by the weight of the filling. In any case the basining of the terrain under each prism means relative thinness and flexibility of the lithosphere and also sufficient weakness in the underlying material to permit its horizontal flow out of the basined sector, and this to a comparatively shallow depth. It is a case of isostatic adjustment, proving weakness in the depths of the sectors involved.

A dozen Pleistocene icecaps, though exerting much less intense pressure on their floors, also basined floors systematically, and again explanation seems clearly to call for belief in isostasy, a relatively thin lithosphere, and an extremely weak interior for our planet.[10] The recoil from the ice load still continues in Scandinavia at a rate measured by R. Witting during his celebrated study of tide-gauge records. Evidently the crust, strong as it is, cannot resist the bending moment due to the differential pressure exerted by the asthenospheric material under Scandinavia. Because the deglaciated tract there is at least 1,200 km. in width, the bending moment for a given central pressure from beneath is much greater than the bending moment due to the excess of the Hidden Range in India or than that in any of the oceanic deeps. If these narrower belts of excess and deficiency of mass are really stable (powerful earthquakes in the belts suggest shearing of the strained crust), a careful comparison of the stress conditions there and in Scandinavia may possibly permit improved estimates of the strength of the earth's crust. It may be noted that the icecaps on the Faroe Islands and Kerguelen Island seem not to have bent or ruptured the crust.

Quite recently R. A. Hirvonen (Veröffentl. finnischen geodät. Institutes, No. 24, 1937) has supplied extraordinarily important data bearing on the strength of the asthenosphere. He reports the Free-air and Bouguer anomalies of gravity at 263 stations, well distributed over the glaciated tract of Finland and a strip of northwestern Russia. The mean Free-air and Bouguer anomalies for all the stations are -7.5 and -16 milligals, respectively, and the mean Hayford anomaly to correspond is about -10 milligals. The mean Hayford anomaly for 40 stations in the central part of the glaciated tract is about -17 milligals. On the highly probable assumption that the Hayford anomalies measure the negative load compelling upwarp of the glaciated tract, computation shows that the maximum stress difference at a depth of 100 km. is of the order of 3 kg. per square centimeter.

Since the central area is still rising at the rate of nearly 1 m. per century, even that amount of stress difference seems unsupportable by the asthenosphere. Thus the strength of this layer appears to be much less than 5 per cent of that assigned by either Barrell or Jeffreys.

The floors of the third kind of veneers, the extensive basaltic plateaus, are basined, but in largest part probably not by mere isostatic adjustment.[11] A veneer of this type, measurable in hundreds of thousands of cubic kilometers, came from the depths, and it is clear that vertical collapse of the rocky floor on which the basaltic lavas came to rest was inevitable. In each major instance that layer has been basined. For example, the surface of the high Cascade Range of mountains was depressed under the thick mass of basalt poured out upon it, and, at the cross section represented by the transverse gorge of the Columbia River, actually depressed below sea level. Similarly the Deccan basalt of India and the huge pile of extrusive basalt in Argentina basined their old crystalline floors, and the plateau basalts of the northern Atlantic basin emerge from the ocean. The same mechanism of collapse seems clear also in the analogous cases of basining under the major lopoliths of basic eruptive material, chiefly basaltic; the Duluth gabbro and the Bushveld basic intrusive are illustrations.

The area occupied by each of these basins is of the same order as the area of the corresponding eruptive body. Hence, either the molten feeder of that body was itself extensive in ground plan, or the material surrounding the feeder was plastic under comparatively small pressures. In either case we must believe that at no great depth under plateau or lopolith the temperature was at the time of eruption high enough to supply an asthenospheric condition.

It is in the highest degree improbable that the underground conditions under geosynclines, geosynclinal prisms or icecaps are peculiar; nor is there any known reason to assume abnormal temperatures in earth sectors when they became capped with plateau basalts or affected by lopolithic injection. Thus in many regions, both continental and oceanic, Nature has provided sampling evidences of the distribution of strength and weakness in the earth—evidences of the reality of lithosphere and asthenosphere as complete earth-shells. Field geology, then, corroborates in principle the picture of the outer earth as drawn by the geophysicists who use the detective methods founded on the use of gravity pendulum and plumb line.

3. *Orogeny.*—All theories of the more complex structures visible in the lithosphere are based on assumptions regarding the nature of the deep interior. Here, however, emphasis will be placed on reasoning in

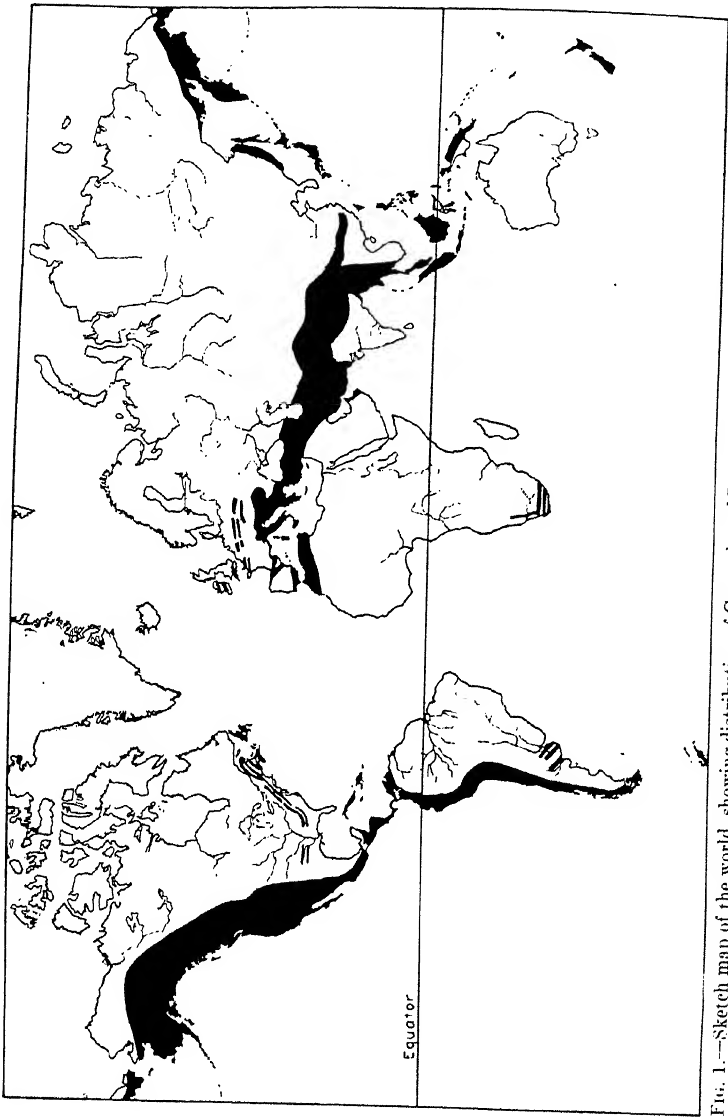

FIG. 1.—Sketch map of the world, showing distribution of Cenozoic and Late-Paleozoic mountain chains. Mercator projection.

reverse, the facts observed at the earth's surface coming first and any clear deductions about the depths being only those that are direct, compelled by the facts, and independent of cosmogonic or other theories.

At the outset it is well not to assume identity of subterranean conditions producing the structural disorder of the Basement Complex with the conditions controlling orogenic movements in Paleozoic and later time. On the other hand, the structures and ground plans of the Caledonide (late Silurian), Hercynide (Permo-Carboniferous), Jurasside (late Jurassic) and Alpide (Cenozoic) chains are so similar that the earth presumably had the same general organization at all four of these revolutionary epochs.

Anything like a full statement of the facts known about mountain chains is naturally out of the question. In principle some of them may be quickly read from Fig. 1, a world map showing in continuous black the approximate ground plans of the post-Cretaceous chains, and with thick, broken lines a few well-mapped stretches of older, largely Hercynide chains. Most of the younger set of chains and, so far as is known, the older set have two loci: bordering the Pacific, and running roughly east and west across North America and Eurasia in mid-latitudes. In addition, the Cape Colony Range and the Paleozoic chain of Argentina and southern Brazil run roughly east and west in mid-latitudes of the southern hemisphere, where, however, there is no conspicuous equivalent of the Alpide system of the north.

A century of intensive research has produced no facts disputing the conclusion of James Hall that the Paleozoic and younger mountain chains are located at, and largely composed of, thick geosynclinal prisms. That the connection is genetic is a matter of inference, but this deduction is so well supported statistically that it has weight approaching that of ascertained fact. The relative weakness of the geosynclinal rocks is one reason, long ago assigned, for their energetic crumpling and thrusting. The second classic reason, however reasonable, is more theoretical, since it postulates weakening of the older lithospheric rocks beneath each prism by rise of the isogeotherms. The quantitative effect of the blanketing and the associated question of the part played by radioactivity in fixing earth temperatures are problems for the geophysicist, and ultimately he may be able to deduce or confirm, from the relation of geosynclinal prism to mountain chain, general truths about the hidden earth-shells.

At least as significant in the search for clues regarding the nature of the outer shells of our planet is a second law, discovered by the field geologists: the original edges of each strongly deformed geosynclinal

prism were brought closer together in the act of mountain making. In other words, the transverse diameter of the prism was shortened. In no case has the amount of the shortening been closely determined. For the Alps the estimates of different authorities vary from a few tens to several hundreds of kilometers. A minimum for the Appalachian and Cordilleran chains is probably much more than 50 km. Whatever the extent of the relative displacement, its reality means that, to some depth yet to be determined, the lithosphere and perhaps also the asthenosphere have undergone horizontal shearing. Field study shows that by shearing scissions, homogeneous solid rock is made heterogeneous and anisotropic. Hence, the orogenic process has here a special bearing on one of the major questions which are appropriately asked during a general inquiry about the interior of the globe. For, specifically, the velocities of the telltale earthquake waves are affected by any lack of isotropy in the constituent rocks of the lithosphere.

The lateral shortening of the orogenic belt is visibly accompanied by the overturning and superposition of rock folds, by thrusts with known amplitudes reaching tens of kilometers, and by rafting thrust-block on thrust-block. These displacements meant thickening of the sial and the formation of sialic roots under the mountain chains. That, in spite of the later erosion, the roots still exist has been indicated by the seismograph, the gravity pendulum and the plumb line, but none of these geophysical evidences is so strong that we can afford to neglect the corroborating facts from field geology. A few of these may be mentioned.

The geosynclinal prisms that were parents to the Appalachian and Rocky Mountain chains rested directly or indirectly upon the Basement Complex. Before the strata of either prism began to be deposited, the surface of the Complex in that belt was relatively flat and not far from sea level. As already noted, the old surface was deeply basined by the growing load of sediment, and it was depressed still more in the act of mountain making. There is no reason to believe that the sialic material older than the prism has been systematically thinned; the horizontal outflow of material, compensating for the vertical displacement, must have been located chiefly in the sima beneath. If, then, the position of the discontinuity between prism and Basement Complex can be determined by the methods of the structural geologist, a minimum thickness for the mountain root is obtained. Manifestly the drawing of the underground contours for the discontinuity is not easy, and any results of such study are at best no more than approximate. Yet, for the purpose of illustrating the general correctness of the root theory, it is worth while to take a few figures from the maps

that R. G. Moss has recently published.[12] He has estimated the thicknesses of the veneers in the United States, thereby giving a rough idea of the depths where a boring machine would be likely to reach the surface of the buried Basement Complex. In the Adirondacks the Basement rocks reach the height of 5,000 ft. (1,500 m.) above sea level. That surface declines to sea level at the Mohawk Valley, is 10,000 ft. (3,000 m) below sea level at the Pennsylvania-New York boundary, and is 20,000 ft. (6,000 m.) below sea level in central Pennsylvania. In Louisiana the old surface is 10,000 to 30,000 ft. (3,000 to 9,000 m.) below sea level. In Colorado it ranges from 12,000 ft. (3,700 m.) above sea level to 13,000 ft. (4,000 m.) below. In general, the described criterion for the existence of mountain roots is met in the Appalachian and Ouachitan chains and in parts of the Rocky Mountain belt. The cause of the local uplift of the old surface of discontinuity far above its original position near sea level is not apparent. Here we encounter another important question which defies an answer until more is known about the conditions deep beneath the visible mountains. However, geologists have conceived no adequate cause for the uplift other than thermal expansion of the lithosphere where the mountain roots are located. If the associated rise of the isotherms has led to actual fusion of the lower parts of these segments of the lithosphere, the vertical expansion entailed by the change of state would be a specially important condition for the rise of the surface. Such expansion, with or without change of state, demands for the whole lithosphere a thermal gradient comparable with that at the surface.

Again, the thicker sedimentary veneers are significant when the character of the upper part of the asthenosphere is in question. The veneers have densities ranging from 2.2 to 2.7, and, according to the testimony of the gravimeter, are close to isostatic equilibrium. This balance has been struck by flow of asthenospheric material away from each sector so covered by accumulated sediments. Unless the density of the displaced, asthenospheric material is considerably less than 3.0, it is hard to understand the approximation to isostatic equilibrium. The difficulty is all the greater if the density below the depth of 40 km. be assumed to be 3.2 to 3.3—a value on which some geophysicists and geologists are basing fundamental conclusions regarding the dynamical history of the globe.

Incidentally, note may be made of the results obtained by Moss for the Great Plains between latitudes 35 and 49° and between longitudes 90 and 100°. In that extensive area the surface of discontinuity averages about 300 m. below sea level—a value of the order expected

if the region is isostatically adjusted for its veneer of Paleozoic and younger sedimentary rocks.

The existence of an extremely weak asthenosphere is implied by every modern explanation of the horizontal component of the pressure responsible for mountain making. After wrestling with the orogenic problem for two centuries, geologists are still comparing speculations about it. Now holding the field are the classic contraction theory, the deep-convection theory and the crust-sliding theory.

The first assumes the horizontal shearing of a comparatively thin *shell of compression* over a thicker *shell of tension*, the relative displacement affecting the whole lithosphere, from each belt of mountains to the antipodes of that belt. Along the zone of scission the shearing stress is not high; hence, there the material must be assumed to have an exceedingly low elastic limit.

The convection theory is obviously based on the postulate of a thin flexible lithosphere and thick asthenosphere; and it is equally manifest that, if downsliding of large segments of the lithosphere has crumpled geosynclinal sediments downstream, those segments must rest on an extremely weak layer beginning at comparatively shallow depth.

The merits of the three theories cannot be discussed here, but it is perhaps worth noting that they are not mutually exclusive. The fact to be emphasized is that the best thought of five generations of geologists regarding orogeny gives good support to the idea that the strong lithosphere is floored by an earth shell with little or no strength.

The utter contrast between the Pacific and Atlantic types of coast line, made finally clear by Eduard Suess, is illustrated in Fig. 1. The Pacific shores are, in general, parallel with, and essentially determined by, the axes of mountain chains, whereas the Atlantic shores truncate such axes. The reason for this difference is in the depths, where are also the keys to the closely related problems of "vanished Old Appalachia" and "vanished Eria" in the region of the north Atlantic, and "vanished Gondwanaland" and "vanished Archihelenis" of the southern hemisphere, and the "vanished Cascadia" of the north Pacific. An adequate theory of the earth's interior must provide conditions that will permit reasonable solutions of these mysteries also. It is significant that the published solutions with any promise at all, whether founded on planetary convection, crust-sliding or wholesale transfer of magmas, agree in postulating an asthenosphere of extreme weakness.

4. *Deformation in the Basement Complex.*—Since the state of the invisible body of our planet at the present time is the theme of this book, it is not necessary even to outline the stratigraphy and general

geology of the Archean terrains. However, as already hinted, there is danger of reaching wrong conclusions about existing conditions of the underground by ignoring certain facts of structure that are peculiar to the Basement Complex. These facts will be briefly stated, and a few of their implications that bear on the nature of the earth's body at the present time will be added.

a. While many extensive sedimentary veneers, which date from the early part of the Paleozoic era or from the late Pre-Cambrian, have escaped important deformation, the bedded rocks of the Basement Complex are crumpled, thrusted or tilted, steep to vertical dips being almost universal.

b. The detailed structures of the Complex are like those in the heart of any recently formed mountain chain, but geologists have found no clear proof that the Archean deformations were determined in general by antecedent thickening of geosynclinal prisms, although this relation may be suspected in a few regions. It looks, rather, as if the intense, continent-wide deformations of the Archean are to be attributed in much greater measure to the other general condition for weakness in the lithosphere, *viz.*, relatively high average temperature. This deduction would mean that during Archean time the lithosphere was notably thinner than in post-Cambrian time.

c. Another contrast: Since the late Archean the eruption of granite on a large scale has been confined to the narrow belts of mountain-building; but granitic eruption affected every part of the Basement Complex, and, as already observed, granite forms the great bulk of the Complex. The colossal size of the successive invasions of this molten matter finds explanation on the double hypothesis that then the lithosphere was thinner than now and also that at intervals of Archean time the lithosphere, wholly or in large fractions, rested on molten granite. This reasonable deduction from the field observations carries with it the idea of a vertical gradient of temperature decidedly steeper than the existing gradient. The corresponding heat would well account for the intense metamorphism of the Archean sediments, a recrystallization which, according to field evidence, was not due merely to extraordinarily deep burial of the rocks concerned.

Nevertheless, the weight of the cover was important at a time when the globe was hotter and was giving off the pegmatite-breeding water at a rate never since equaled. Then the recrystallization took place under the vertical, non-hydrostatic pressure of the covers, and to this type of change geologists have given the name *load metamorphism*. The evidence for its efficiency in Pre-Cambrian times seems clear from the petrographic study of the occasional flat-lying beds of the Base-

ment Complex and also from the petrography of veneering sediments of late Pre-Cambrian age.

Thus the relations of the invading granites and the phenomenon of widespread load metamorphism confirm the naïve conclusion of the early geologists that the Archean earth was specially hot. The idea is in a sense banal, but no geologist, geophysicist, or cosmogonist has yet found its full quantitative meaning for the problem of the earth's temperature in the twentieth century.

d. The recognition of load metamorphism in the Basement Complex leads to the question whether the same type of recrystallization has taken place in the lower part of the lithosphere as it exists today. If so, the rocks there are likely to be characterized by more or less horizontal schistosity or foliation, analogous to the textures prevailing in the Archean terrains. Therewith we again face the possibility that these deep rocks are anisotropic to a degree which may conceivably affect the velocities of earthquake waves at the corresponding levels.

e. Another fact about the deformations recorded in the Basement Complex may turn out to have much meaning for the student of the earth's internal temperature at the present time. The several unconformities, erosion breaks, registered in the Complex and also many other items in its history show that the total time taken for its development was at least as long as that from the dawn of the Paleozoic onward; more probably the ratio exceeds 2 to 1, a conclusion that seems well supported by studies of the radioactive elements (see Table 3, to follow). For enormous stretches of time, therefore, the deforming forces were highly efficient, the lithosphere seems to have been specially thin and magmas invaded it on a scale never since equaled. The rule of such conditions through a large part of a billion years obviously means that the supply of heat from the interior was decidedly greater than it is now. In explanation there is a natural appeal to the original high temperature of the young earth, but the conditions of the Archean would become still more intelligible if it could be proved that the radioactivity of the rocks was then considerably greater than at present. One of the heating agents is "potassium forty," K_{40}, which has a half-life much shorter than uranium or thorium and yet still continues to heat the rocks.[13] During early Archean time this isotope was delivering heat about twice as fast as at present. The effect on temperature was in this case not great, but K_{40} illustrates a question of principle, a question whose answer may conceivably involve a veritable revolution in thought about the earth's internal temperatures both during the Archean and at present. For expert students of radioactivity are considering the possibility that the young earth was heated

by radioactive elements with short lives, only those with long lives having actually been discovered in the rocks. If such energetic elements had lives of a few millions of years or a few scores of millions of years and were distributed in the depths of the earth, the temperatures of the Archean and of the twentieth century would have to be higher than those computed from the present rate of the furnace heating.

The idea represents a truly appalling complication in the problem of this book. If it could be proved essentially correct, it would be indeed difficult to state the thermal conditions of the globe during the first half-billion years of its life. But the point here to be emphasized is that any uncertainty on that subject tends to prevent worth-while computation of the thermal condition at the present time. Actual estimates of this have been based on assumptions as to the distribution of heat in the young earth, and in practice an initial thermal gradient, approximately fixed by the temperature of freezing of igneous rock at terrestrial pressures, has been postulated. Since neither convection nor conduction is likely to have removed the excess of heat from the deep interior, some of this excess would have to be allowed for in calculating the present internal temperature.

f. Although some suggestions have been offered about the conditions of deformation in the Basement Complex, the cause of the deformation is a subject that has been left, and must long remain, wide open. Here, then, we have a question, not an explanatory fact; but the question seems worth asking, even on the present occasion. It is now clear that the crumpling and upending along the axes of the post-Cambrian chains of mountains are results of enforced lateral movement of the sial toward the respective axes and that these movements meant local thickening, concentration, of the sial in the orogenic belts. In view of the continent-wide crumpling and upending visible in the Basement Complex, is it possible that the present restriction of nearly all of the sial to one hemisphere is due to a Pre-Cambrian concentration of a thinner layer of crystallized sial which at an earlier time had covered the whole of the young earth? Geology knows no harder question, but, whether answerable now or not, it represents a fundamental fact that must not be lost sight of when the effort is made to picture the earth's body as it now exists, for manifestly good reasoning about the present constitution and heat content of the planet has to allow for the horizontal segregation of the sial. And the method of the segregation itself needs scrutiny, if the full bearing of that ancient, revolutionary change on the conditions of our own day is to be discerned.

UNGROUNDED HYPOTHESES OF DEFORMATION

The literature of geology contains many expressions of belief in two kinds of major deformation of the lithosphere, neither of which is of the geosynclinal or orogenic kind. An example is found in the widely held explanation of coral reefs, atolls and barriers, on the assumption of strong and continued subsidence of much of the intertropical belt. Another example is found in the explanation of the deeply submerged "canyons," cut in the outer slopes of the continental shelves.[14] Some geologists have thought these canyons to have been cut by rivers, after the shelves were uplifted many thousands of feet and before the same shelves sank back to practically their original levels.

If these two hypotheses were correct, a good theory of the earth's body would have to provide for the physical conditions—forces and potential energies—that would permit the described gigantic warpings of the lithosphere. However, both hypotheses are fast losing adherents. The good charts published during the last 40 years prove the essential, long-continued stability of the lithosphere in most of the belt occupied by atoll and barrier. And the discovery of deep-sea canyons in every ocean and on the periphery of every continent has shown the fallacy of the foregoing explanation of the canyons. Both atoll and canyon seem much better understood as by-products of changes wrought in the sea by Pleistocene glaciation.[15] In any case the student of the earth's interior need feel no responsibility for these two imagined types of distortion of the globe.

Geophysicists have set up two other hypotheses involving still greater distortions, with important consequences for the theory of the earth's interior. It is well to note that neither of these, the third and fourth hypotheses, is well supported by the observations of field geologists.

According to one of these speculations, connected particularly with the names of D. Kreichgauer,[16] A. Wegener[17] and E. Haarmann,[18] the earth's axis of rotation and its crust have undergone relative displacements through arcs measurable in tens of degrees. Wegener's evidence for such drastic changes has been derived chiefly from the distribution of ancient icecaps, the traces of which are now at low latitudes. He assumed that the glacial sheets with the known great dimensions could have been formed only in high latitudes. In view of our profound ignorance of the cause of even Pleistocene glaciation, many geologists are reluctant to accept this criterion for proving extensive migration of the poles. Wegener's argument is specially doubtful when based on the glaciation of the last million years. More-

over, the geologist has quite failed to find any of the systematic deformations that the lithosphere should have suffered if the poles had migrated so far and so recently. Although no wise geologist would deny the possibility of polar wandering in the past, he cannot regard great relative motion of the kind as a fact that should exercise an important control over thought about the internal conditions.

The second geophysical hypothesis here to be noted is that, if the equator of the geoid or sea-level figure of the earth is an ellipse, there would have to be considerable stress difference and presumably corresponding strength throughout the whole of the silicate shell above the iron core. Yet it is quite conceivable that the earth's triaxiality, if real, is due essentially to the asymmetry of the centrosphere, assumed to be pressure-solid and strong. In that case the triaxial figure would be stable, even if the asthenosphere has zero strength. Incidentally we note that, if the thick lower part of the silicate shell is a comparatively strong solid, we can better understand the high average viscosity of the planet as computed by N. A. Haskell.[19]

It may be added that triaxiality for the geoidal figure of the earth has recently been deduced by both Hirvonen[20] and Heiskanen,[21] using more and better data than those available for previous determinations of the figure. Hence it seems quite possible that geological proof of extreme weakness of the asthenosphere may be found to demand the assumption of considerable strength for a thick layer still deeper within the earth. Such extension of deductive reasoning from facts observed at the surface of the globe is, however, a subject beyond the scope of this chapter.

DATA FROM THE ERUPTIVE ROCKS

The most promising cosmogonic theories lead to the conclusion that for a brief time, or for brief times, the earth was fluid at the surface. Without assuming at least one stage of thorough mobility, it is practically impossible to explain the increase of intrinsic (chemically determined) density from sial, through sima, to the so-called *iron core* of the planet, or to explain many principal facts of geology. On the other hand, this debt of geologist to cosmogonist may be in part repaid by suggestions of the geologist as to the way in which a molten earth would be expected to organize itself. In fact, the new science of petrology has suggestions. How they have been developed out of a host of field observations is a long story. Space fails for more than a mere enumeration of the principles involved.

1. About 700 varieties of igneous rocks are named in the handbooks, but these messengers from the depths do not represent an equal

number of primary melts below the surface of the lithosphere. Many of the species, *e.g.*, the andesites and the less voluminous quartz diabases, trachytes, syenites, anorthosites, many pyroxenites and peridotites, the magmatic sulphide rocks, and chromitite are derivatives of basaltic liquid, either pure or contaminated during reaction of that liquid with the rocks or enclosed volatile matter of the lithosphere. As yet, none of the species just named, except basalt, has been proved to have any other origin. This is particularly worthy of note in the case of peridotite, which includes dunite.

Typical basalt itself, eruptible and at intervals erupted since an early Pre-Cambrian date, has all the characteristics of a melt that may be called *primary* in the sense that it has not been contaminated by material from the sial. We have seen that enormous masses of basalt have flooded the lithosphere in both oceanic and continental sectors, and its composition is nearly uniform, wherever and whenever erupted. These and other facts have led to the conception of a world-circling shell of basalt, at no great depth, a theoretical view that is fast winning its way. However, among petrologists there is a sharp difference of opinion as to the state of the basaltic shell, one group assuming it to be crystalline throughout except for local "pockets" of liquid and another group assuming a universal, vitreous sublayer of basaltic composition.

Still more divergent are the explanations offered for the huge masses of granitic and allied magmas locally intruded into the lithosphere during post-Cambrian time. Nevertheless, all of the more promising speculations on the matter place the sources of these melts no deeper than the base of the basaltic layer. Apparently, therefore, theories about the younger granites will not contribute much to the problem of conditions below a depth of a few tens of kilometers from the earth's surface.

2. Magmatic temperatures as high as 1200°C. have been found at basaltic volcanoes. No temperature so high as 1400°C. is demonstrated. From observations in field and laboratory it appears that important subcooling of natural magmas has never occurred except at the surface, where the rapid radiation of heat does give the vitreous state represented by obsidian and tachylite. If, then, a rapid outflow of basalt reaches the surface at a temperature of 1200°C., a nearly identical temperature must be attributed to the source region in the basaltic shell. And the hypothesis that each source region is a localized pocket in an essentially holocrystalline earth, rather than a continuous earth-shell of vitreous basalt, seems irreconcilable with the theory of a cooling earth.

Nor does the pocket hypothesis agree well with the distribution of the active, dormant and recently extinguished volcanoes.[22] Their well-known alinements in the famous Circle of Fire around the Pacific and along the Tertiary orogenic zones across Central America and Europe may be recalled by a glance at Fig. 1, where the run of the genetically connected mountain structures is shown. Other pronounced alinements are represented by the Hawaiian, Samoan, Society, and Azores groups of islands and the row of recent submarine eruptions in mid-Atlantic under the equator. These thousands of recent eruptions took place either where the lithosphere was orogenically ruptured or in loci of dominant and widespread tension in the lithosphere. The geographical relations described are hard to understand at all on the pocket hypothesis but, on the other hand, directly suggest the tapping of a vitreous, eruptible *couche* below the lithosphere and world-circling. We have already deduced a similar origin for the plateau basalts.

It appears, too, that the much debated genesis of the giant batholithic intrusives is shorn of a deal of mystery if the same assumption of a vitreous, basaltic substratum is made.

3. The densities of the holocrystalline rocks and the equivalent glasses at different temperatures and pressures are of fundamental importance in certain aspects of the present inquiry. Some of the more reliable data may be summarized.

Table 2 refers to materials at room temperature (20°C.). The crystallized types carry from 0.5 to 1.5 per cent of volatile matter by weight, and the percentages of volatiles in the rhyolite glass and tachylite are nearly the same. On the other hand, the glasses studied by Douglas[23]—marked J.A.D.—and by Day, Sosman and Hostetter[24]—marked D., S., H.—had lost their original volatiles during the process of melting. In these seven cases the percentage decrease of density due to the change of state, when no volatile matter is lost, would be slightly greater than as shown in the last column of the table. In the first and third columns the figures in parentheses indicate the respective numbers of determinations averaged.[25]

For standard rocks the changes in the coefficients of thermal expansion and of volume compressibility have not been measured with all desirable accuracy, but some deductions from experiments already made seem clear. (1) Down to the bottom of the sial the densities of holocrystalline and vitreous granite, quartz monzonite, granodiorite and other quartz-rich rocks, if mineralogically unchanged, would retain almost perfectly their respective values at the surface

of the earth. (2) The densities of normal gabbro, diabase, dolerite and basalt and their tachylitic equivalents, if mineralogically unchanged, would show, with increasing depth, a slow decrease, reaching a maximum of about 0.03 not far from the 20-km. level, and would then slowly increase. At the 50-km. level the densities would be about 0.03 greater than under the conditions at the surface. (3) The densi-

TABLE 2

Type of rock	Density	Equivalent glass	Density	Percentage decrease of density due to change of state
Granite (155)	2.667	Rhyolite glass (15)	2.370	11.1
Granite (1, J.A.D.)	2.656	Granite glass (J.A.D.)	2.446	7.9
Granite (1, J.A.D.)	2.630	Granite glass (J.A.D.)	2.376	9.66
Syenite (1, J.A.D.)	2.724	Syenite glass (J.A.D.)	2.560	6.02
Tonalite (1, J.A.D.)	2.765	Tonalite glass (J.A.D.)	2.575	6.87
Diorite (1, J.A.D.)	2.880	Diorite glass (J.A.D.)	2.710	5.90
Dolerite (1, J.A.D.)	2.925	Dolerite glass (J.A.D.)	2.800	4.27
Diabase (1, D., S., H.)	2.975	Diabase glass (D., S., H.)	2.761	7.19
Diabase (40, fresh)	2.980	Tachylite (11)	2.772	7.0
Gabbro (27)	2.976	Tachylite (11)	2.772	6.85
Dunite (1, fresh)	3.289			
Lherzolite (1)	3.33			
Wehrlite (1)	3.37			
Pyroxenite (8)	3.231			
Anorthosite (10)	2.734			
Gneiss, mica schist	2.6–3.1			
Sandstone, argillite	2.2–2.8			
Limestone	2.6–2.8			

For granite containing the normal amount of volatile matter, which is retained also in the vitreous state, a good round figure for the decrease of density in changing state is, at 20°C., 10 per cent. The corresponding round figure for gabbro, diabase, or dolerite is 6 per cent.

ties of the intermediate types of rock would also change little if transferred from the surface to the depth of 50 km.

It appears, therefore, that the contrasts of density between holocrystalline phase and equivalent vitreous phase are not greatly changed by pressure and temperature, down to a depth considerably exceeding 50 km. This fact is significant in connection with theories involving the differentiation of magmas and earth-shells, as well as what may be called solid-liquid convection.

4. A large number of the intrusions called *sills, laccoliths* and *lopoliths* and even some lava flows at the earth's surface show, in the passage from top to bottom, systematic change of chemical and mineralogical composition. In each of many instances the change was evidently brought about by gravity, heavier fractions developed in the melt having sunk and lighter fractions having risen before final solidification. To the process as a whole the name *gravitative differentiation* has been given, and the fractions that sheared their way downward or upward may be conveniently called *units of differentiation.*[26] These units include sulphides which, being only to a limited degree miscible in the prevailing molten silicate and also of high density, separate as sinking globules. Some petrologists believe that metallic iron also has limited miscibility, with settling-out as a consequence. That silicate fractions behave similarly has never been proved, and most authorities are of opinion that the silicate units of differentiation are crystals and the complementary liquid fractions. Both crystallization and remelting of any common igneous rock are progressive. The early crystals of a freezing magma are generally denser than the residual liquid and tend to sink. The residual liquid itself, being less dense than the original liquid, tends to rise through any of this original liquid that exists in the magmatic body at the levels where the local crystallization takes place. The first liquid formed during remelting is of relatively low density and tends to rise, whereas the residual solid matter with its relatively higher density tends to sink. There are excellent reasons to believe that all of these conditions controlled the gravitative differentiation registered in the visible intrusive bodies. The relative importance of fractional crystallization and fractional remelting is, however, a subject of controversy.

5. There is no good evidence that ordinary thermal convection has effectively stirred any eruptive mass. On the other hand, vertical stirring of lava columns by differential vesiculation and by the loss of magmatic gas to the atmosphere is clearly shown at volcanic vents like Kilauea and Mokuaweoweo (Mauna Loa). This stirring, often continuous for periods of days or months, may be distinguished under the name *gas-liquid convection.* It is one kind of *two-phase convection.*[27] Another kind, *solid-liquid convection,* though theoretical and outside the realm of direct observation, may have had incomparably greater importance in organizing the outer shells of the earth.[28] At surface outcrops and in mines and tunnels, blocks of the invaded rock are seen enclosed in granitic batholiths, and in many cases it is clear that the blocks had sunk through the batholithic magma to the visible levels, the liquid correspondingly rising. Lord Kelvin imagined that the original

crusting of the earth took place by such *magmatic stoping* on a gigantic scale; he supposed flakes of the incipient crust to founder into the earth's body, until the consequent chilling of the body permitted the formation of a coherent, stable crust. If the internal shells were originally superheated or, through radioactivity, became superheated, such sinking flakes would undergo progressive melting, and the secondary liquids would rise toward the earth's surface, to be refrozen. Then the process would be repeated. The relative movements of solid and liquid mean complete overturn of material, *i.e.*, a kind of convection. This speculation regarding one way in which the lithosphere originated cannot be easily tested by the field geologist, but none the less it represents a vital question for the student of the earth's interior.

6. Because gravity controlled the differentiation of so many floored injections, it is natural to refer the original development of the earth-shells to a similar cause. This suspicion is strengthened by a few outstanding facts: the manifest superposition of the sial on the basaltic sima in continental sectors of the globe; the values of the earth's mean density and of its moment of inertia, which prove limited thickness for the granitic and basaltic layers together and the existence of a third, deeper and much thicker shell of silicate matter which is intrinsically denser than basalt; and the high density of the earth's core. Unless the visible, differentiated intrusives are quite misleading, it also seems reasonable to assume the material at the surface of the infant earth to have been initially denser and more "basic" than basalt. For the nature of that material we look to the stony meteorites and are almost inevitably led to question whether the moon, with mean density of 3.3, does not represent the earth's superficial matter at some early stage of the evolution.

7. Less speculative is a conclusion derived from the known general arrangement of the earth-shells, however and whenever these were individualized. If thermal convection has ever stirred the earth to great depth, it must have been of the step-by-step or "tandem" variety.[29] That is, any deep shell could not be convectively overturned until the overlying shell or shells had already been overturned by chilling against outer space. Single-step convection, affecting the whole silicate shell down to the assumed iron core, seems to have early become impossible because of the gravitative differentiation of the planetary magma as a whole.

8. The existence of dikes, particularly such dike swarms as those of the British Isles, prove brittleness for at least the upper part of the lithosphere. That its rock is brittle even when heated to temperatures of many hundreds of degrees centigrade is shown by the shatter zones surrounding, and the angularity of blocks included in, batholithic

masses—blocks torn from roof and wall and incorporated by these hot melts. For these and other reasons it appears unsafe to postulate important plastic stretching of the lithosphere, at geosynclines, under the newer ocean basins or elsewhere.

9. Finally, a word about a general fact declared by the petrologist's microscope; the dominant minerals constituting the accessible rocks are packed with actual, or "realized," cleavages and are correspondingly porous. There is reason to believe that some of these openings are not closed in specimen rocks even by an all-sided pressure of 20,000 atmospheres, unless the temperature should be raised to a point higher than any yet accompanying such high pressure on rock in the laboratory. On the other hand, realized cleavage can hardly exist in rock minerals at the pressures and temperatures ruling 30 km. or more below the earth's surface. Hence adequate correction for this contrast between accessible and deep, inaccessible rocks of the same composition should be attempted, if the laboratory measurements of either rigidity or compressibility of rocks are used for detecting the character of the lower part of the lithosphere by means of the velocities of seismic waves.

TIME SCALE FOR THE DEVELOPMENT OF INTERNAL CONDITIONS

We are concerned with an evolution from the cosmogonic stage of the earth's history to our own epoch. We therefore need to know how much time has been available for such processes as radiation of heat, radioactive generation of heat, accumulation of sedimentary rocks, the making of mountain structures, igneous action and isostatic adjustment. For convenience of reference, Table 3 has been prepared. Its data have been compiled from A. Knopf's summary statement in the elaborate report on The Age of the Earth (Nat. Research Council Bull. 80, Washington, 1931); and from C. Schuchert's estimates of elapsed time from the facts of stratigraphic geology, combined with the ages of certain formations as derived from the lead-uranium ratio in contained minerals.[30]

In addition Urry's measurements of the ages of 22 meteorites by the helium-radium method may be noted. These ages vary from 100 million to 2,800 million years. Twelve meteorites gave ages between 1,450 and 2,800 millions of years and four had a range from 2,300 to 2,800 millions. All of these values are provisional (see Ref. 29).

SUMMARY OF CONCLUSIONS

In review, the facts of field geology, bearing on the problem of the earth's interior, may be listed. Some of the most important are of negative character.

1. *Facts to be considered when seismic-wave velocities are used for diagnosing the outer earth-shells.*

a. The geologist clearly sees that both load metamorphism and orogenic shearing have developed anisotropy in the visible part of the lithosphere, and may suspect an even greater effect of the kind in the lower half of the lithosphere. Yet the geologist cannot tell the degree of failure of isotropy and thus help the geophysical detective to estimate its relation to wave velocities.

TABLE 3

AGES OF ROCK FORMATIONS, IN YEARS

Era and period	Time elapsed since beginning of era or period, millions of years: Measured by uranium method	Measured by the Schuchert method
Cenozoic era:		
Pleistocene		
Pliocene		
Miocene	....	37
Oligocene	< 34	
Eocene	70	60
Mesozoic era:		
Cretaceous	....	120
Jurassic	>123	150
Triassic	....	175
Paleozoic era:		
Permian	>220	210
Pennsylvanian	....	255
Mississippian	278	290
Devonian	....	330
Silurian	<349	355
Ordovician	> >371	395
Canadian	....	415
Ozarkian	....	440
Upper and Middle Cambrian	....	490
Lower Cambrian	....	515
Pre-Cambrian eras:		
Keweenawan		
Keewatin (lava)		
Archean (various stages)	1,000–1,770 (Holmes)	

b. The geologist finds realized, or actual, cleavage dominating in the minerals of the accessible rocks and has good reason to doubt that open cleavages exist at depths exceeding 20 km. In any case the geophysicist should inquire as to the possible effect of this more perfect bonding of crystalline material on its rigidity and compressibility and therefore on the wave velocities expected in that material.

c. The geologist finds that under high stress crystallized basaltic material is not stable as ordinary gabbro or diabase but may be stable as garnet-bearing gabbro or as amphibolite or as an allied type of rock. On the other hand, he has not discovered a single large body of eclogite with the chemical composition of basalt.

d. Many observed masses of peridotite, including dunite, are fractional derivatives of basalt. There is no certain field evidence that dunite of different origin exists in the earth, though some kind of peridotitic material probably does underlie the basalt.

e. The geologist has found no large body of tachylite among the thousands of mapped basaltic intrusives, and his field experience tends to negate the hypothesis that a tachylitic shell can constitute any important sublayer of the lithosphere.

f. Field geology amply confirms the deduction of the seismologists that the dominating material of the sial (upper sublayer of the continental lithosphere) is granite. Hence it is appropriately called the *granitic sublayer*.

2. *Facts to be considered in discussion of the existing thermal gradient.*

a. The great length of geological time has been proved by the methods of stratigraphic geology, the result of which is of the same order as that obtained by the lead-uranium methods.

b. That for at least a large fraction of 1,000 million years, in Pre-Cambrian time, the thermal gradient was much steeper than that now ruling has been proved by the field geologist. This fact casts doubt on the hypothesis that the earth has been cooling steadily since the epoch of an initial crusting. The events registered in the Basement Complexes of the continents become more intelligible if it is understood that the radioactive generation of heat was many times more rapid in the young earth than it is at present.

c. Molten basalt has been erupted at intervals from an early Pre-Cambrian date to the twentieth century and at points well distributed over both continental and oceanic sectors of the globe. The rate of increase of temperature with increase of depth at the present time must be such as to provide an adequate source for this magma of world-wide distribution.

3. *Facts bearing on the chemical composition of the outer earth-shells.*

a. The geologist has found no important body of rock which may be considered as planetesimal material, and he has no evidence compelling belief that the earth has been crystalline at the surface ever since it attained its present mass.

b. On the other hand, the relation of sial and sima *inter se* and also their relation to the mean density of the planet strongly suggest original fluidity and gravitative differentiation, analogous to that observed in many intrusive sheets. These sheets show in succession, from above downward, granite, intermediate rocks, and rock of basaltic composition. There are many technical reasons for assuming a similar succession in the lithosphere.

c. Field geology does not support the hypothesis that there is a syenitic layer under the ooze of the sea floor.

4. *Facts bearing on the problem of the distribution of strength.*

a. The endurance of granite, gabbro, diabase and other constituents of sial and sima at and near the earth's surface, though stressed for millions of years.

b. The relative thinness of the strong layer (lithosphere), as shown by open geosynclines, geosynclinal prisms, icecaps and general isostasy.

c. Extreme weakness below the lithosphere to the depth of some hundreds of kilometers, as shown by the displacements represented in glaciated regions, in general isostatic adjustments and in orogeny. On the other hand, geology knows no facts forbidding belief that there may be considerable strength in a thick earth-shell underlying the asthenosphere.

5. No geologist has been able to reconcile his world of field observations with the speculation that the earth is essentially *crystalline* down to the iron core or with the related speculation that a thick shell of crystallized *dunite* begins anywhere near the 40-km. level below the surface of the globe. Correct picturing of the earth's interior must give an *earth model that works*, that is, explains the facts of dynamical geology. Geophysics and geology must go hand in hand.

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CHAPTER IV

ELASTIC PROPERTIES OF MATERIALS OF THE EARTH'S CRUST

L. H. ADAMS

(*Director, Geophysical Laboratory, Carnegie Institution of Washington*)

A knowledge of the elastic behavior of rocks is required for the solution of many problems in geophysics. Of especial importance is the correlation of the velocity of earthquake waves with the velocities calculated from laboratory measurements on the elasticity of typical rocks. Such comparisons provide the simplest and most direct evidence concerning the constitution of those parts of the earth that are inaccessible for direct observation. The following is a summary of the pertinent data in this field. From the numerous experimental results that now exist there have been selected those that are especially useful in connection with investigations of the earth's interior. No attempt has been made to include all measurements, since many of these have merely a historical interest.

COMPRESSIBILITY OF MINERALS

Definitions.—Compressibility is usually defined as the relative change in volume per unit change in pressure referred to an arbitrary initial pressure; *i.e.*,

$$\beta = \frac{1}{V_0}\frac{\partial V}{\partial P}, \tag{1}$$

in which β denotes the compressibility, V the volume of the sample in question at the pressure P, and V_0 the volume at the arbitrary initial pressure. It is here understood that the temperature, chemical composition and physical state of the sample remain constant and that the pressure to which it is subjected is purely hydrostatic. The unit of pressure employed here is the bar, which is defined as 10^6 dynes per square centimeter. In latitude 40° 1 kg. per square centimeter is equal to 0.9806 bar. The bulk modulus is the reciprocal of the compressibility; *i.e.*,

$$k = \frac{1}{\beta}. \tag{2}$$

Compressibility may also be defined as $\frac{(\partial V/\partial P)}{V}$, or as $\partial v/\partial P$, in which v is the volume per gram of the material, but these forms are less generally useful in considerations relating to the elasticity of rocks and minerals. Numerical values of β are here expressed as reciprocal bars.

The rigidity μ may be defined as the ratio of the shearing stress (or its change) per unit area to the shearing deformation (or its corresponding change). That is,

$$\mu = \frac{\partial P_s}{\partial s}, \tag{3}$$

in which P_s is the shearing stress per unit area and s equals $\partial l/\partial x$, x being the distance along one axis as reference and l the corresponding displacement in the y direction, the stresses being in plane parallel to the z direction. The shear $\partial l/\partial x$ is equal to the change in angle between adjacent sides of a square that has been distorted into a rhombus and is also equal to twice the relative change in length of either diagonal of the original square.

Young's modulus E is defined as the relative change in length l of a rod of uniform cross section, caused by a given change in stress per unit area, in the direction of the axis of the rod. That is,

$$E = \frac{1}{l_0}\frac{\partial l}{\partial P}. \tag{4}$$

Poisson's ratio σ is equal to the ratio of the lateral extension to the longitudinal contraction of a rod subjected to a thrust along its axis. The relation between k, μ and σ is as follows:

$$\mu = k\frac{3(1 - 2\sigma)}{2(1 + \sigma)}. \tag{5}$$

Apparatus and methods.—Measurements of the compressibility of minerals have been made by several methods. Voigt[11] used an indirect method, determining various elastic constants by bending and twisting a crystal and then measuring the deformation produced by a given stress. It appears that this method can yield useful results for some minerals, but the difficulty of measuring the small effects produced by the application of the maximum twisting moment and bending moment that the crystal can withstand is very great, and consequently this method is less valuable than the other methods that have been used in more recent years.

Richards and Jones[9] and Madelung and Fuchs[8] measured the compressibility of minerals under hydrostatic pressure, using a piezometer.

Satisfactory precision can be obtained by this method, but the results obtained through its use have been limited to pressures of no more than a few hundred bars.

The piston-displacement method was found by Adams, Williamson and Johnston[4] to be a satisfactory one for determining the compressibility of minerals and other solids and has been used at the Geophysical Laboratory for a number of years in the determination of the volume change under pressure of various liquids and solids. This method was apparently first used by Cowper and Tammann.[34] Later, in an improved form, it was employed by Parsons and Cook[35] for measuring the compression of liquids at pressures up to 6,000 bars. Further improvement in accuracy and a considerable extension of the pressure range are due to Bridgman,[33] who used the piston-displacement method at pressures up to 20,000 bars—principally, however, for obtaining the compressibilities of liquids.

In essence, the principle of the method is as follows. A leakproof piston is forced into a heavy-walled steel cylinder which contains the specimen completely surrounded by a liquid that will not freeze or become unduly viscous at the highest pressure to be employed. Pressures are measured, to within 1 bar or better, by means of an electrical pressure gage which depends on the change of electrical resistance of a coil of manganin wire immersed in the liquid. The movement of the piston corresponding to a given change in pressure, or the equivalent volume change, is composed of three principal parts—diminution in volume of the sample itself under pressure, the volume change of the liquid by which the pressure is transmitted and the effect due to the distortion of the bomb and piston packing. In order to eliminate the last two of these three effects, another experiment is performed in which an equal volume of a substance of known compressibility is substituted for the material under investigation. In effect, the quantity that is measured by this method is the difference in compressibility of the sample and of a reference material. Iron is commonly used as the reference substance; its compressibility at pressures up to 12,000 bars has been measured with high precision by Bridgman.[36,28] At 25° the volume change of iron at various pressures (in bars) is represented by the equation

$$-10^6\frac{\Delta V}{V_0} = 0.598P + 0.0243 \times 10^{-4}P^2. \qquad (6)$$

Very high precision in compressibility measurements has been obtained by Bridgman[36] by an apparatus in which the change of length of a solid specimen under hydrostatic pressure is measured. The

specimen in the form of a rod is placed in a bomb to which fluid under pressure is supplied. One end of the specimen is held in a fixed position within the bomb, and the other end, being free to move, actuates a contact sliding along a short length of fine wire. By a suitable measurement of electrical resistance the contraction of the specimen relative to the container is determined. Finally, after making the necessary corrections, the change in length of the specimen at each of several pressures becomes known. Since the longitudinal extension of the container is difficult to determine with high accuracy, the customary procedure is to calibrate the apparatus by means of a specimen of known compressibility.

This method, when used for nonisotropic materials, has the disadvantage of requiring separate measurements on two or more specimens cut along different directions in the material. Even for apparently isotropic substances, such as glass, there may be a sensible difference in the elastic behavior for different directions owing to lack of annealing or to inhomogeneities with respect to chemical composition. This effect is likely to occur in pseudo-isotropic crystalline aggregates. Although it is now known that many aggregates, including coarse-grained granites, do not show much variation in the elastic behavior for various directions, nevertheless, in the use of the linear method for determining compressibilities, the presence or absence of effective isotropy needs to be investigated for each kind of material under consideration. Provided that suitable precautions are observed, this linear method yields results of satisfactory precision even with substances whose compressibility is less than 10^{-6}. The inherent sensitivity of the method is much greater for solid substances than that of any other method that has been used at pressures of several thousand bars.

For isotropic substances there is a very simple relation between the relative change of length and the relative change of volume. If the *linear* contraction $-\frac{\Delta l}{l_0}$ is represented by the equation

$$-\frac{\Delta l}{l_0} = AP - BP^2, \tag{7}$$

A and B being constants, then the relative diminution in *volume*, *i.e.*, the compression, will be represented by the equation

$$-\frac{\Delta V}{V_0} = 3AP - 3(A^2 + B)P^2, \tag{8}$$

the unimportant terms of order higher than P^2 being neglected.

TABLE 4

COMPRESSIBILITY (β) OF MINERALS IN RECIPROCAL BARS

Composition	Name	Temperature, °C.	Pressure, bars	$10^6\beta$	Pressure range	$-10^4 \frac{1}{\beta_0}\frac{\partial\beta}{\partial P}$	Ref.
	Stibnite	0	125	1.50	50–200	....	8
	Bismuthite	0	125	3.32	50–200	....	8
	Diamond	25	7,000	0.18	1–12,000	....	3, 32
	Quartz	25	1	2.7	1–12,000	0.17	3, 6, 8, 11
ZrO_2	Rutile	25	1	0.50	1–12,000	0.09	8, 13
PbS	Zircon	0	125	0.86	50–200	....	8
$PbSO_4$	Galena	25	1	1.88	1–12,000	0.09	6, 8
$PbCO_3$	Anglesite	0	125	1.94	50–200	....	8
ZnO	Cerussite	0	125	1.91	50–200	....	8
ZnS	Zincite	0	125	0.78	50–200	....	8
ZnS	Sphalerite	25	1	1.30	1–9,000	0.04	6, 8
	Wurtzite	0	125	1.36	50–200	....	8
$MnCO_3$	Argentite	25	1	2.7	1–9,000	0.03	6, 8
Fe_2O_3	Rhodochrosite	0	125	1.3	50–200	....	8
Fe_3O_4	Hematite	0	125	0.60	50–200	....	8
FeS_2	Magnetite	0	125	0.55	1–12,000	0.08	6, 8
FeS_2	Pyrite	25	1	0.70	1–12,000	0.07	3, 6, 8
$FeAsS$	Marcasite	0	125	0.82	50–200	....	8
$FeCO_3$	Arsenopyrite	0	125	0.99	50–200	....	8
Fe_2SiO_4	Siderite	0	125	1.00	50–200	....	8
$FeTiO_3$	Fayalite	25	7,000	0.90	2,000–12,000	....	17
$CuFeS_2$	Ilmenite	0	125	0.56	50–200	....	8
$CoAsS$	Chalcopyrite	0	125	1.29	50–200	....	8
Al_2O_3	Cobaltite	25	1	0.77	1–12,000	0.09	6, 8
$3FeO \cdot Al_2O_3 \cdot 3SiO_2$	Corundum	25	1	0.35	1–12,000	....	8, 19
$3BeO \cdot Al_2O_3 \cdot 6SiO_2$	Almandite	25	7,000	0.57	1–12,000	....	14
MgO	Beryl	25	1	0.56	1–12,000	0.09	8, 11, 13
Mg_2SiO_4	Periclase	25	1	0.62	1–12,000	0.08	8, 15
$3MgO \cdot Al_2O_3 \cdot 3SiO_2$	Olivine	25	7,000	0.79	1–12,000	....	17, 24
CaF_2	Pyrope	25	1	0.55	1–12,000	0.09	13
$CaSO_4$	Fluorite	25	1	1.21	1–12,000	0.13	6, 8, 11
$CaSO_4 \cdot 2H_2O$	Anhydrite	0	125	1.84	50–200	....	8
$Ca_3P_2O_8 \cdot CaF_2$	Gypsum	0	125	2.50	50–200	....	8
$CaCO_3$	Apatite	25	1	1.09	1–12,000	0.10	13
$CaCO_3$	Calcite	25	1	1.36	1–12,000	0.06	4, 6, 8, 11
$3CaO \cdot Fe_2O_3 \cdot 3SiO_2$	Aragonite	0	125	1.55	50–200	....	8
$3CaO \cdot Al_2O_3 \cdot 3SiO_2$	Andradite	25	1	0.68	1–12,000	0.07	13
$CaCO_3 \cdot MgCO_3$	Grossularite	25	7,000	0.60	1–12,000	....	14
$CaSiO_3 \cdot MgSiO_3$	Dolomite	0	125	1.22	50–200	....	8
$SrSO_4$	Diopside	25	7,000	1.60	1–12,000	....	3
$SrCO_3$	Celestite	25	1	1.56	1–12,000	....	6, 8
$BaSO_4$	Strontianite	0	125	1.75	50–200	....	8
$BaCO_3$	Barite	25	1	1.77	1–12,000	0.14	8, 13
Al(S	Witherite	0	125	2.03	50–200	....	8
aCl	Spodumene	25	1	0.72	1–12,000	0.08	6, 13
	Halite	25	1	4.18	1–12,000	0.20	4, 7, 8, 9, 10, 11
$Na_2O \cdot Al_2O_3 \cdot 4SiO_2$	Jadeite	25	1	0.75	1–12,000	....	14
Na_2SO_4	Thenardite	25	1	2.37	1–12,000	0.20	2
KCl	Sylvite	25	1	5.65	1–12,000	0.27	8, 9, 10
$K_2O \cdot Al_2O_3 \cdot 6SiO_2$	Orthoclase	25	1	2.13	1–12,000	0.15	13
$KCl \cdot 2Na_2CO_3 \cdot 9Na_2SO_4$	Hanksite	25	1	2.45	1–12,000	0.20	13

Minerals of complex composition

Composition	Name	Temperature, °C.	Pressure, bars	$10^6\beta$	Pressure range	$-10^4 \frac{1}{\beta_0}\frac{\partial\beta}{\partial P}$	Ref.
$En_{88}Fs_1$	Enstatite	25	7,000	1.00	1–12,000	....	3
$En\ _0Fs_1$	Hypersthene	25	7,000	0.98	1–12,000	....	3
A $_8An$	Oligoclase	25	1	1.74	1–12,000	0.12	3
$_8An$	Labradorite	25	1	1.50	1–12,000	0.13	3, 14
Ab_9	Microcline	25	1	1.9	1–12,000	0.15	3
	*Phlogopite mica	25	1	2.34	1–12,000	0.17	3
	Actinolite	25	7,000	1.29	1–12,000	....	3
	Augite	25	7,000	1.01	1–12,000	....	3
	Tourmaline	25	1	0.82	1–12,000	0.07	8, 13
	Topaz	25	1	0.61	1–12,000	0.08	11, 13
	Opal (Mexican)	0	125	6.1	50–200	....	
	†Jeffersonite	25	1	0.91	1–12,000	0.12	13

En is written for $MgSiO_3$; Fs for $FeSiO_3$; Ab for $Na_2O \cdot Al_2O_3 \cdot 6SiO_2$; Or for $K_2O \cdot Al_2O_3 \cdot 6SiO_2$; An for $CaO \cdot Al_2O_3 \cdot 2SiO_2$.

* Essentially $R_3Mg_3Al(SiO_4)_3$ where R is H, K, Mg or F.

Essentially $CaO \cdot 3(Mg,Fe)O \cdot 4SiO_2$.

Essentially $CaSiO_3 \cdot (Mg,Fe)SiO_3$ (diopside) with Al_2O_3 and Fe_2O_3.

Essentially $H_9Al_3(BOH)_2Si_4O_{19}$ with Fe, Mg, Ca, F, Na and K.

Essentially $Al(F,OH)_2SiO_4$.

† Essentially $CaSiO_3 \cdot (Mg,Zn,Fe,Mg)SiO_3$ with small amounts of Al_2O_3.

Results.—In Table 4 the most important of the measurements on the compressibility of naturally occurring minerals have been collected. This table gives for each mineral the composition, the mineral name and the value of $10^6\beta$ at the temperature and pressure indicated respectively in columns 3 and 4. The temperature as given is in most instances that at which measurements were made; in some instances an interpolation or a short extrapolation has been made. In column 6 is shown the variation of compressibility with pressure. It should be noted that $\frac{10^4(\partial\beta/\partial P)}{\beta_0}$ is a convenient unit for this variation, since it is the fractional decrease in β for a pressure change of 10,000 bars.

A study of the measurements on this table shows, first, that there is a surprisingly small range in the values of this quantity for various minerals. As pointed out by Adams and Williamson,[3] a rough approximation for the compressibilities of silicate minerals (pure SiO_2 being excepted) may be calculated from certain values assigned to the constituent oxides. These values are as follows: SiO_2, 1.4; Al_2O_3, 0.8; FeO and Fe_2O_3, 0.5; MgO, 0.7; CaO, 0.8; K_2O and Na_2O, 6.0. The computation is made by multiplying the number of molecules (by weight) of each oxide by the corresponding factor and dividing the sum of these products by the total number of molecules. This calculation will ordinarily give the compressibility with an accuracy of 10 or 15 per cent (usually much closer) and is of interest because it shows the predominant effect of soda and potash in increasing the compressibility of minerals.

COMPRESSIBILITY OF ROCKS

Methods.—Measurements on the compressibility of rocks may be carried out by the same methods that have been applied to minerals. It should be noted that piezometers of the type used by Richards and Jones[9] and by Madelung and Fuchs[8] have not been used for rocks. This is probably because these methods were not employed at high pressures. As is now known, a considerable pressure range is required for an adequate determination of the characteristic volume change of rocks under pressure.

Adams and Williamson[3] called attention to the two procedures by which the compressibility of a porous material, such as a rock surrounded by a liquid under pressure, could be determined. If the liquid is directly in contact with the rock, it will penetrate the pores and will act directly on the individual grains of the solids. The compressibility would then be dependent mainly on the separate compressibilities of the grains. If, however, the solid is covered by a thin

impervious coating such as tin foil, the volume decrease when hydrostatic pressure is applied would include the effect of packing together the various homogeneous parts of the solid. Measurements on both covered and uncovered specimens of typical rocks were carried out. At high pressures the differences in the values obtained by the two methods were not large. The effect is of greatest importance at pressures less than 1,000 bars. It is obvious that experiments on enclosed specimens have a more direct bearing on the elastic behavior of rocks at considerable depths below the surface of the earth.

The measurement of the compressibility of rocks by the linear method, used with much success in connection with single crystals or a portion of a single crystal, has been a difficult problem. It has now been shown by Birch and Bancroft[37] that in order to apply the linear method successfully to materials like rocks the specimens must be enclosed. The linear contraction of unenclosed porous rocks depends almost entirely upon the skeleton of the less compressible crystals. Another important advantage in enclosing the sample is that the application of pressure then tends to improve the elastic properties of the rock and to minimize irregularities, especially at low pressures, whereas the application of pressure by means of a liquid that is allowed

TABLE 5
COMPRESSIBILITY OF ROCKS
(Nonhydrostatic pressure. Indirect method. Ref. 1)

Description	Pressure, bars	$10^6\beta$	Pressure range
Westerly granite	300	3.2	1– 600
Baveno granite	300	3.2	1– 600
Peterhead granite	300	3.0	1– 600
Lily Lake granite	300	3.2	1– 600
Quincy granite	300	3.4	1– 600
Stanstead granite	300	3.7	1– 600
Montreal nepheline-syenite	300	2.3	1– 600
Sudbury diabase	300	1.36	1– 600
New Glasgow gabbro	300	1.5	1– 600
New Glasgow anorthosite	500	1.7	1–1,000
Mount Johnson essexite	300	2.1	1– 600
Black Belgian marble	300	1.7	1– 600
Carrara marble	300	2.4	1– 600
Vermont marble	300	2.7	1– 400
Tennessee marble	300	2.4	1– 600
Montreal limestone	300	2.3	1– 400
Ohio sandstone	140	8.	1– 270
[Plate glass]	300	2.23	1– 600

to penetrate the pores eventually causes a notable deterioration of the specimen.

In past years there were numerous attempts to measure the compressibility of rocks by indirect methods involving nonhydrostatic stresses. The first successful effort in this direction was made by F. D. Adams and E. G. Coker.[1] From measurements of Young's modulus and Poisson's ratio, they calculated various elastic constants such as bulk modulus and compressibility, assuming that the specimens were essentially isotropic. A summary of their results is shown in Table 5. The stress-strain curves of typical rocks obtained by these investigators show an appreciable amount of hysteresis. This is a common effect for crystalline aggregates subjected to nonhydrostatic stresses. The pressure range here is obviously limited by the crushing strength of the material.

Measurements of the compressibility of rocks by the piston-displacement method are shown in Table 6. This "cubic" method, although not so sensitive as the linear method, yields satisfactory results with porous as well as nonporous materials and has sufficient accuracy to determine the change with pressure of the compressibility for all rocks except those with very low compressibility. The measurements for the various granites shown on the table pertain to enclosed specimens; for other rocks not much difference between enclosed and unenclosed samples is observed by the cubic method, and measurements of these two varieties for rocks other than granites are listed indiscriminately.

In Table 7 are given various measurements of the compressibility of rocks by the linear method, the applied pressure being hydrostatic. Most of the measurements are at low pressures—less than 1,000 bars. The measurements by Zisman[16] were carried out with both covered specimens—the results shown in the table—and with uncovered specimens. For reasons already given, the results obtained by the linear method for materials that have an appreciable porosity do not necessarily give the true values of the compressibility. Zisman's measurements provide a striking demonstration of the abnormal value of compressibility that can be obtained for unenclosed specimens by the linear method, especially at very low pressures. A critical discussion of Zisman's results has been given by Goranson.[21]

The recent measurements by Birch and Bancroft[37] were carried out with covered specimens. Their values for compressibilities at 4,000 bars and 30°C. are shown below in Table 8. It is interesting to compare the values obtained for Maryland diabase by the cubic method and by the linear method carried out on covered specimens of rock

TABLE 6

COMPRESSIBILITY OF ROCKS

(Hydrostatic pressure. Direct measurement of volume change. Pressure range, 2,000 to 12,000)

Description	Pressure, bars	$10^6\beta$	$-10^4 \frac{1}{\beta_0}\frac{\partial\beta}{\partial P}$	Ref.
Westerly granite........	2,000	2.12	0.19	3
Washington granite.....	2,000	2.23	0.24	3
Stone Mountain granite	2,000	2.06	0.17	3
Sudbury diabase........	2,000	1.37	0.11	3
Palisade diabase.......	2,000	1.54	0.2	3, 14
Maryland diabase.....	2,000	1.23	0.16	14
New Glasgow gabbro...	2,000	1.34	0.2	3
Whin Sill diabase......	2,000	1.70	0.3	14
New Jersey basalt......	2,000	2.4	0.4	3
Balsam Gap dunite.....	7,000	0.79	....	17, 24
Serpentine............	2,000	1.79	0.31	3
Colorado marble.......	7,000	1.37	....	3
Obsidian..............	2,000	2.82	0.07	2, 3
Tachylite (basalt glass).	7,000	1.45	....	24
[Plate glass]...........	2,000	2.22	0.10	3

TABLE 7

COMPRESSIBILITY OF ROCKS

(Hydrostatic pressure. Volume change, determined from change of length)

Description	Pressure, bars	$10^6\beta$	Pressure, bars	$10^6\beta$	Ref.
Quartzite sandstone...............	300	3.55	600	3.15	16
Quincy granite (from 235 ft.).......	300	3.18	600	2.55	16
Rockport granite..................	300	3.39	600	2.71	16
French Creek norite...............	300	2.64	600	1.69	16
Sudbury norite...................	300	1.75	600	1.68	16
Olivine diabase...................	300	1.46	600	1.33	16
Vermont marble...................	300	2.09	600	1.53	16
Limestone........................	300	2.60	600	2.40	16
Solenhofen limestone (not covered)..	6,000	1.39	...	....	5
Dolomite.........................	300	1.89	600	1.51	16
Lipari obsidian....................	300	3.01	600	3.01	16
Lipari obsidian....................	5,000	2.7	...	....	28
Ascension Island obsidian..........	5,000	2.61	...	....	12
Diabase glass.....................	5,000	1.62	...	....	20, 28

from the same locality. The values at 4,000 bars are respectively 1.19×10^{-6} and 1.16×10^{-6}. In view of the fact that these are compressibilities at a definite pressure and not the mean compressi-

bilities over the experimental pressure ranges, the agreement should be considered excellent.

Relation between the compressibility of a rock and that of the constituent minerals.—Upon comparing the compressibilities of various rocks with the compressibilities of the minerals of which the rocks are composed, it may be seen at once that the acidic rocks have a high compressibility and consist of minerals of high compressibility, whereas the basic rocks have a much lower compressibility and consist of less compressible minerals. This generalization was put in quantitative form by Adams and Williamson[3] who showed that, except at very low pressure, the compressibility of a rock can be calculated with satisfactory accuracy if the separate compressibilities of its various minerals are known. In this calculation the average is taken with respect to volume rather than weight. At a pressure of 10,000 bars the agreement between the calculated and the directly measured compressibilities was found to be of the order of 2 per cent except for rocks that obviously had been altered. In such instances the alteration of the rock may have reduced the compactness of the structure and consequently produced an abnormally high compressibility. At very low pressures the agreement between the observed compressibility and that calculated from the mineral content is not so satisfactory, but at pressures as low as 2,000 bars the average agreement between "observed" and "calculated" values of β was about 5 per cent for fresh rocks. For a diabase from Palisade, N. J., containing a considerable amount of saussuritized feldspar, on the other hand, there was a discrepancy of 25 per cent between the two values. The circumstance that at moderately high pressures the compressibility of a fresh holocrystalline rock is an additive function of the compressibilities of its minerals in spite of considerable porosity is of great practical importance; it naturally suggests that, in view of some of the difficulties inherent in the direct determination of the compressibility of rocks, it may possibly be preferable to determine this property, especially for coarse-grained rocks, indirectly from measurements on the compressibility of its constituent minerals. In this connection it is interesting to note the correspondence, shown by Adams,[25] between the compressibilities and pressure coefficients of compressibility for crystalline compounds.

The further consequence of this generalization is that at pressures of about 2,000 bars the compressibility of various rocks of a given general type is nearly the same. This is illustrated by Fig. 2, which is taken from the paper by Adams and Gibson.[14] The width of the shaded areas indicates the variation that may be expected for ordinary

rocks within a given class. This diagram shows the abnormal increase of compressibility at very low pressures and the relatively small variations at pressures above 1,000 or 2,000 bars.

Porosity.—All naturally occurring crystalline rocks have an appreciable porosity. Many of the difficulties encountered in the measurement of the elastic properties of rocks are dependent directly or indirectly upon the porosity of the structure. Pores may be divided into two classes: closed pores or cavities separated from one another

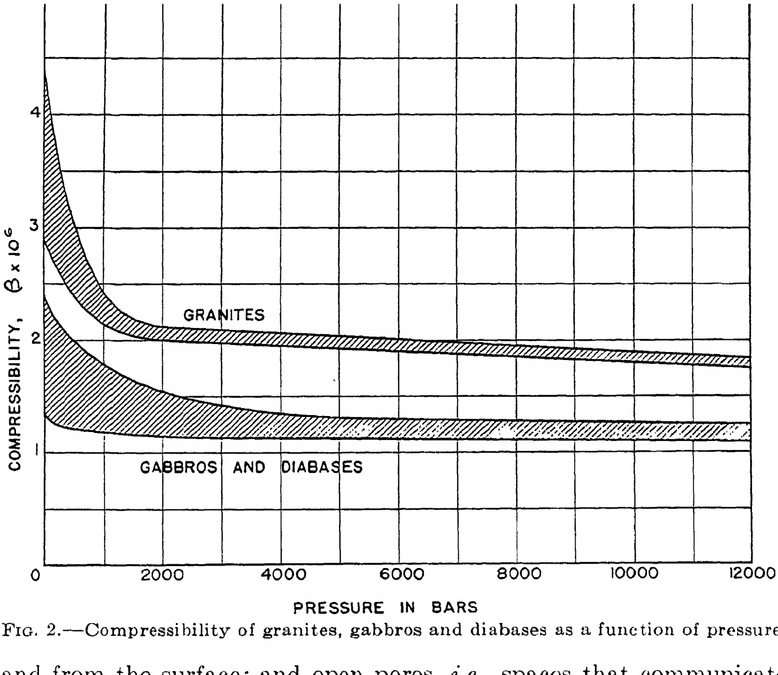

FIG. 2.—Compressibility of granites, gabbros and diabases as a function of pressure.

and from the surface; and open pores, *i.e.*, spaces that communicate with the surface of the given specimen. Closed pores would not by themselves have an important influence on the compressibility of a rock. It can be shown that if a solid contains separated spherical cavities amounting in volume to 1 per cent of the total volume, the compressibility would be approximately 2 per cent greater than that of a similar solid having no porosity. The gabbroic rocks ordinarily have a porosity whose order of magnitude is 0.1 per cent of the total volume. The porosity of granitic rocks is notably higher but seldom exceeds 1 per cent. The open spaces that contribute the porosity are

made up of irregular thin cracks separating the various crystal grains. In the measurement of the compressibility of rocks subjected to hydrostatic pressure, the porosity in the case of unenclosed specimens should have little effect on the compressibility, provided that the cubic method is used and that the pressure fluid is able to penetrate freely into the interstices. As mentioned above, the application of the linear method to unenclosed porous rocks produces marked difficulties.

Granitic rocks as a class are relatively porous and show at low pressures the most pronounced abnormality in elastic properties. This effect for various kinds of materials of similar porosity is more common with structures consisting of an assemblage of two or more crystalline components. As pointed out by Adams and Williamson,[3] an important underlying cause of the peculiarities exhibited by granitic rocks is the circumstance that such rocks are combinations of hard and soft minerals or, more precisely, of highly compressible and less compressible mineral grains. Investigations by Birch and Bancroft[37] have made it clear that the elastic behavior of rocks becomes simple only by the application of sufficient pressure so that the mineral grains become tightly locked. We have, then, at least two effects contributing, especially at low pressures, to the anomalous elastic behavior of ordinary rocks: (1) a complicated set of stress conditions with respect to the individual mineral grains, and (2) an initial looseness of structure that may be produced in a variety of ways. It has been suggested that the porosity and lack of compactness of granites may be due to the inversion in quartz, which takes place at 575° with a large change of volume.

Effect of temperature.—There have been few satisfactory estimates of the change of compressibility with temperature for rocks. Birch and Law[28] and Birch and Dow[29] succeeded in measuring the compressibility of several materials, including a sample of Vinal Haven diabase, at pressures up to 10,000 bars and at temperatures up to 500°. It is now known that the results for the diabase, having been obtained with an unenclosed specimen, do not represent the true compressibility of the material. It is probable that the temperature coefficient of compressibility obtained by them also needs revision. Measurements by Birch and Law on obsidian and diabase glass are not subject to this uncertainty, which applies only to crystal aggregates. For obsidian it was found that the compressibility increases at first with increasing temperature, attaining a maximum at about 150°. Above this temperature the compressibility decreases. Diabase glass shows a similar behavior, the maximum compressibility being at about 175°. Silica glass decreases steadily throughout the experimental temperature

range, and other commercial glasses show varying behavior with respect to the temperature coefficient of compressibility.

At present, our total knowledge concerning the effect of temperature on the compressibility of rocks is very small indeed. Fortunately the method and the apparatus for determining this important quantity have been developed, and additional valuable information on this subject will probably be forthcoming in the near future.

Comparative compressibility of vitreous and crystalline materials.—Apparently the compressibility of a liquid is always greater than the compressibility of a crystalline solid of the same composition and at the same temperature and pressure. If glasses are considered as undercooled liquids (which is true with respect to some, if not all, considerations), it is reasonable to expect that the compressibility of a vitreous rock would be greater than that of a crystalline rock of the same composition. It should be noted in this connection that the densities of glasses and liquids are, with few exceptions, lower than the densities of the corresponding crystals (*i.e.*, the specific volume is higher). There is, however, no quantitative parallelism between the specific volume and compressibility. Obsidian, with 10 per cent greater volume, has 40 per cent greater compressibility. On the other hand, silica glass,[2,12,28,28,29] with 20 per cent greater volume than quartz, has 15 per cent greater compressibility. The compressibility of tachylite is about 25 per cent greater than chemically similar diabase.

RIGIDITY AND POISSON'S RATIO

Methods and results.—Whereas the concept, compressibility or bulk modulus, can be readily employed with nonisotropic substances or aggregates as well as with liquids or purely isotropic solids, other simple elastic properties, such as rigidity and Poisson's ratio, are directly applicable only in connection with isotropic materials such as glass or with pseudo-isotropic solids such as crystalline aggregates. There are two principal classes of methods for measuring the rigidity of rocks. Measurements by Adams and Coker[1] already referred to in connection with the determination of the compressibility of rocks yield information also concerning rigidity and Poisson's ratio. More recently, Zisman[38] used a similar static and indirect method, and Ide[22,23] has compared the elastic constants obtained by static and by dynamic methods.

Such measurements, though valuable for a number of purposes, are not completely applicable to the problem of the elastic behavior of material at great depths within the earth: (1) on account of the relatively small range of stress, and (2) on account of the important

simplification in elastic behavior caused by subjecting the specimens to hydrostatic pressure. Birch[27] and Birch and Bancroft[37] employed a dynamic method whereby they measured the velocity of torsional waves in various rocks at temperatures from 30 to 100° and under hydrostatic pressures as high as 4,000 bars. The specimen in the form of a cylinder is set into oscillation by means of a torque applied to one end. Measurement of the frequency of vibration at which resonance occurs gave by a simple calculation the velocity of the torsional waves and thence the rigidity of the material. By combining these results with the separately determined compressibilities, Poisson's ratio was calculated. The results are shown in Table 8. The values of β for quartzitic sandstone and for marble in this table were obtained by other investigators. These measurements of the rigidity of rocks are of the highest importance. They give us for the first time an accurate notion of the rigidity of various typical rocks under conditions existing at considerable depths below the surface of the earth.

Variation with temperature and pressure.—For crystalline rocks the rigidity increases with increasing pressure, at first rapidly, then more slowly. As shown by Bridgman,[26] the rigidity of metals usually

TABLE 8

ELASTIC CONSTANTS AND WAVE VELOCITIES FOR VARIOUS ROCKS AT 4,000 BARS AND 30°C.

(From Birch and Bancroft[37])

Description	$10^6\beta$			V_p, km./sec.	V_s km./sec.
Quartzitic sandstone	$(24.0) \cdot 10^{-1}$	$4.28 \cdot 10^{-1}$	0.118	6.08	4.00
Solenhofen limestone	21.4	2.47	0.276	5.54	3.08
Vermont marble	(13.9)	3.33	0.299	6.51	3.49
Granite:					
Quincy 1	19.2	3.45	0.229	6.08	3.61
Rockport	18.5	3.36	0.243	6.24	3.59
Syenite, Ontario	16.9	3.15	0.274	6.05	3.36
Norite, Sudbury 2	14.4	3.81	0.268	6.49	3.65
Diabase:					
Vinal Haven	11.7	4.46	0.277	6.97	3.88
Maryland	11.6	4.42	0.281	6.96	3.83
Gabbro:					
Mellen	11.4	4.00	0.302		3.71
French Creek	11.3	4.80	0.270	7.15	3.98
Pyroxenite:					
Hypersthenite	9.6	6.86	0.230	7.83	4.58
Bronzitite	8.9	6.80	0.249	7.86	4.55
Dunite	8.3	6.84	0.262	8.05	4.57

increases with pressure. Here again we have an illustration of the abnormal behavior of rocks at pressures of less than 1,000 bars. The mean pressure coefficient of rigidity in the pressure range from 1,000 to 2,000 bars lies within the range 4 to 80 p.p.m. per bar for the various rocks, whereas in the pressure range 3,000 to 4,000 bars the pressure coefficient varies from 1×10^{-6} to 20×10^{-6} for the same rocks.

On the other hand, rigidity normally decreases with increasing temperature. For the various rocks investigated the temperature coefficient of rigidity covers the range of values from about 100 to about 300 p.p.m. per degree. Whereas the pressure coefficient is on the average about the same for acidic as for basic rocks, the temperature coefficient increases steadily with increasing basisity. Obsidian and pyrex glass were found to be abnormal, the rigidity increasing with increasing temperature—at least within the temperature range 30 to 100°.

WAVE VELOCITY AS DETERMINED BY MEASUREMENTS OF ELASTICITY

The velocity V_p of longitudinal waves and the velocity V_s of transverse waves transmitted through a material are related to the density and the elastic constants of the material, according to the following equations:

$$V_p = \sqrt{\frac{k + \frac{4\mu}{3}}{\rho}} \tag{9}$$

$$V_s = \sqrt{\frac{\mu}{\rho}}, \tag{10}$$

ρ being the density. Strictly speaking, the compressibility (reciprocal of k) employed in this equation should be the adiabatic compressibility and not the isothermal compressibility, which is the one ordinarily measured. The difference between these two compressibilities for rocks is not much greater than the error of measurement. It is customary, therefore, to neglect this difference and to use the isothermal compressibility without correction. If Poisson's ratio were constant for all solids, k would bear a fixed relation to μ. For example, if σ equals 0.27, μ equals $0.543k$. Adams and Williamson,[3] having noticed that seismological data showed that at depths greater than about 50 km. σ is very nearly constant and equal to 0.27, used their compressibility measurements to calculate the rigidity of rocks and also the

velocities of longitudinal and transverse waves. For typical granites they obtained 5.6 km. per second for V_p at a pressure of 2,000 bars and at ordinary temperatures. It is now known that, at relatively shallow depths, σ is less than 0.27; at 10 km. depth, σ, calculated from recent seismological data, is found to be 0.20; at 30 km. depth, 0.25.

A small error in the assumed value of σ will have a considerable effect upon the wave velocities calculated from compressibility measurements alone. For example, changing σ from 0.27 to 0.26 will increase V_p by 1 per cent and V_s by 2.5 per cent, the compressibility remaining the same. As pointed out by Birch and Bancroft,[37] Voigt's measurements on the elasticity of quartz indicate that σ should be 0.07 for a pure quartz aggregate. The average value of σ obtained by Birch and Bancroft for granites is 0.236. For diabase and also for norite and syenite σ was found to differ very little from the value 0.275. The lower value of σ found for granites is undoubtedly connected with the content of quartz in these rocks.

From the measurements by Birch and Bancroft[37] on two different granites the longitudinal velocities are 6.1 and 6.2 km. per second respectively at 4,000 bars and 30°. The temperature coefficient of compressibility for granites has not yet been determined, but it seems improbable that correcting the elastic constants to the average temperature appropriate for the granitic layer of the earth would lower the velocities mentioned above by more than a few per cent. The velocities obtained by Birch and Bancroft for granite are notably higher than those obtained by Adams and Williamson. The difference is due (1) to the lower value of σ compared with that assumed by Adams and Williamson, and (2) to the somewhat lower value of β found by Birch and Bancroft for granites. From seismological data V_p for the layer 10 or 15 km. in thickness lying immediately below the superficial layer of sedimentary rocks is found to be 5.5 km. per second On the other hand, careful observations by Leet[30,31,39] and by Wood and Richter[40] of the velocity of longitudinal waves from quarry blasts in granitic regions indicate the velocity to be 6.0 km. per second. We have here a puzzling situation. The general effect of pressure is to increase the wave velocity, especially with granites at moderate pressures, but the velocity at considerable depths appears to be higher than that near the surface. It is also worthy of note that the value of σ obtained from seismologic data (0.20) is significantly lower than that observed by Birch and Bancroft for granites. In view of the large number of considerations that point to a granitic shell immediately beneath the sedimentary layer, it is difficult to give up the notion of a granitic shell despite the fact that it does not seem easy to reconcile the

velocities determined in various ways. A similar discrepancy exists for the transverse waves. From seismologic data, V_s is 3.36 km. per second, whereas from quarry blasts it has been found to be 3.5. Birch and Bancroft by laboratory measurement find V_s for granites to be equal to 3.53 km. per second at 3,000 bars and 290°, which represents approximately the average temperature and pressure in the granite layer. These authorities suggest the possibility that the granite samples hitherto investigated do not fully represent all granitic rocks and that waves traveling in the "granitic layer" may pass through a considerable amount of metamorphic rock.

In the "intermediate" layer, about 25 km. in thickness and lying immediately below the granitic layer, V_p from seismologic data is approximately 6.3 km. per second. From the early measurements by Adams and Williamson, V_p was calculated to be 6.9. Later measurements by Adams and Gibson[14] on four samples of diabase gave values ranging from 6.6 to 7.2 at 10,000 bars and room temperature. These results may be compared with the determinations by Birch and Bancroft, who found 6.96 and 6.97 for two samples of diabase at 4,000 bars and 30°. For V_s corrected to 8,000 bars and 720°C. Birch and Bancroft obtained 3.7. Comparing these results with 6.3 and 3.65, which are, respectively, the values of V_p and V_s from seismologic data for the intermediate layer, we may note that, although the results for V_s are consistent with a diabasic composition for the intermediate layer, the results for V_p do not match, unless a very high temperature coefficient for V_p is assumed.

Recent estimates of V_p and V_s from seismologic data for the material just below the intermediate layer are 7.8 km. per second and 4.35 km. per second. From measurements on dunite (an olivine rock), Adams and Gibson[24] concluded that V_p in this material at 10,000 bars and room temperature is 8.2 km. per second. The result obtained by Birch and Bancroft is 8.1, also at room temperature but at 4,000 bars. By extrapolation of their measurements of V_s to 12,000 bars and 960° these authors obtain 4.1 km. per second for V_s. On the basis of present knowledge there does not appear to be a complete correspondence between the seismologic velocities in the layer under consideration and the velocities determined by laboratory measurement for either pyroxenite or dunite, which, except for the unusual rock eclogite, are the only types of rock having wave velocities high enough to meet the seismologic requirements. It appears, however, that there are no measurements that conflict seriously with the original conclusion by Adams and Williamson that the material lying below the crust of the earth consists of ultrabasic rock.

Velocities of earthquake waves in a vitreous layer of basaltic composition can be inferred from the compressibility measurements by Adams and Gibson[14] on tachylite and by Birch and Law[28] on a glass made by fusing diabase. As pointed out by Birch and Bancroft, the assumption that σ is equal to 0.23 leads to values of 6.8 and 4.0 km. per second for V_p and V_s, respectively, in the tachylite and of 6.5 and 3.5 in the "diabase glass."

It is evident that attempts at identification of the material at various depths within the earth in terms of known rock types continue to offer many interesting problems. Encouraging progress has been made, and some puzzling questions already have been answered. Additional measurements are needed with many other samples and over as wide a temperature range as possible.

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CHAPTER V

THE CRUST OF THE EARTH AND ITS RELATION TO THE INTERIOR*

H. S. Washington

(*Geophysical Laboratory, Carnegie Institution of Washington*)

Much is known about the crust of the earth and very little about the part that lies far below the surface. What information we have concerning the interior depends to a large extent on observations of the composition and physical state of the crust. Conversely, in order to comprehend fully the constitution of the crust and its relations to the earth as a whole, one must know something of what lies beneath it, for there is good reason to believe that the crust and the interior are related genetically, so that an understanding of one must aid in an understanding of the other.

The relation of the crust to the interior aroused the interest of the earliest philosophers, who speculated as to what supported the known or visible "earth," whether the broad shoulders of a giant, an elephant standing on a tortoise, or "the waters under the earth." In more recent times the general belief, accepted as an assumedly necessary consequence of the Kant-Laplace nebular hypothesis and upheld to within a few years ago, was that the earth consists of a thin, solid crust of cooled material overlying and floating on a liquid, still molten interior. In favor of this view, indeed in proof of it, were held to be the existence of volcanoes and the efflux of lava at the surface, the presence of masses of intrusive igneous rock in the crust, the existence of hot springs and the increase of temperature with increase of depth, the gradients being such that by linear continuation all rocks would necessarily be molten at a few miles below the surface. Such a structure was also thought necessary to permit the profound movements and dislocations that the crust has undergone. One of the last upholders of this doctrine was Osmond Fisher, an Englishman,[1] who estimated the solid crust to be about 25 miles thick.

* A manuscript bearing this title was prepared by Dr. Washington shortly before his death. Subsequently the material was condensed and rearranged by L. H. Adams. Further revision with respect to various details and a final editing have been carried out in order to make the section conform with the other parts of this volume.

But this view of a thin crust resting on a mobile liquid interior is now almost or quite universally abandoned, although the term crust, which was based on this theory, still survives to denote the outermost solid layer of the earth. The chief reason for this change of view is that an earth so constituted, with a relatively thin crust, could not withstand the disruptive forces of the lunar and solar tidal pulls or the distortional stresses set up by precession and polar nutation. Hopkins, in 1839, was the first to advance the adverse argument based on precession and nutation, and Lord Kelvin, in 1862, advanced that of the tides.* Roche,[2] also, showed mathematically that such a structure was incompatible with the amount of polar flattening and with that of precession. Another adverse argument that has been brought up more recently is that, inasmuch as rocks expand on melting, the enormous pressures in the interior must keep them solid, even if at a temperature very much above the melting point (see section on Crystallinity). Another reason is that the transverse earthquake waves, which will be mentioned again later, are not transmissible through a liquid. Other objections to the hypothesis of a thin crust on a truly liquid interior have been adduced, and the general outcome is that this concept of the structure of the earth is now discarded by geologists and geophysicists, although it has still a strong popular hold.

It is now the general belief that the earth is essentially rigid, at least down to the core, and that it is not homogeneous radially but that there are important variations in the character, composition and state of the material from the surface to the center. Some of these variations will be considered in the succeeding pages.

Although some sort of progressive change downward is now believed in by all geophysicists, there is uncertainty and considerable diversity of opinion as to the details of these changes, such as, for example, the number and the relative thickness of the various shells and whether they grade continuously from one to the other or are sharply separated by discontinuities.

The terms in use to denote the various divisions are not very satisfactory, as they are unsystematic and unrelated, and some of them are ambiguous. Thus, the term crust has been used in more than one sense or with different limits; *lithosphere* has been applied to the stony portion of the interior as well as to the peripheral crust; *centrosphere* is commonly used to denote all the matter below the outer crust, although it should etymologically apply only to the core or nucleus at the center.

There would be some advantage in using the terms *kentrosphere* for the core or nucleus, *mesosphere* for the intermediate shell and *perisphere*

* These arguments are discussed by Fisher (Ref. 1, Chap. 3).

for the uppermost layer or crust. The perisphere might conveniently be divided into the *epiperisphere* (the outer accessible part) and the *hypoperisphere* (the part inaccessible to direct observation). However, in conformity with more common usage and with the terminology adopted for other sections in this volume, the terms core, mantle and continental layers will be adopted here for the three important divisions of the earth. Instead of continental layers, the term crust will be used when the condition of crystallinity is to be emphasized.

THE CRUST OF THE EARTH

The term *crust*, as has been said, dates back to the time when the earth was thought to consist of a thin solid shell resting on a molten, liquid interior and is now used to denote the outermost, peripheral, certainly solid, portion of the globe. This is composed mainly of igneous rocks, with a very thin veneer of sedimentary rocks; locally both igneous and sedimentary rocks have been changed by pressure, and other agencies, to metamorphic rocks. In the consideration of the constitution of the earth as a whole, the sedimentary and the metamorphic rocks are of negligible importance, however great they may loom in general geology.

For the outer 10 miles of the earth's crust, Clarke[3] has estimated the rock composition of the crust to be about as follows: igneous rocks, 95.0 per cent; shales, 4.0 per cent; sandstones, 0.75 per cent; limestones, 0.25 per cent. Such masses as coal beds or salt and ore deposits are of little significance in connection with a study of the chemistry of the crust as a whole (though their presence is of great importance in other respects), and the superficial layer of loose soil is absolutely negligible.

Igneous rocks are those that have solidified from a state of fusion, or rather liquidity, as the term *fusion* implies a previous solid condition. The liquid matter, which eventually solidifies as a rock, is called technically the *magma*—a term that is in frequent use in petrology. The magma comes up from below; from what depth we do not know, though there is some reason for thinking that in many instances the places of origin are not very deep. Nor do we know whether it arises from the melting of portions of the earth that are actually solid but potentially liquid on relief of pressure or whether it is, in general, derived from "reservoirs" of liquid magma.

The igneous magma may be compared, as it usually is, to a complex solution of salts in water. This idea, which was first suggested by Bunsen in 1861, is of great importance and has been very fruitful in our study of the origin, formation and characters of igneous rocks.

It is of interest to note that, although about 1,000 different minerals are known, yet the number of the different kinds that compose by far the great majority of igneous rocks—certainly over 99 per cent by weight of these—is very small. Indeed, the really important and essential igneous rock-forming minerals number only about a dozen, *viz.:* quartz, silicon dioxide; the feldspars, silicates of alumina and potash, soda or lime, including the potassic orthoclase, the sodic albite and the calcic anorthite, with isomorphous mixtures of these; the pyroxenes, metasilicates of calcium, magnesium and iron, with aluminum or sodium in some cases; the amphiboles, in chemical composition much like the pyroxenes but differing in crystal form and otherwise; the micas, aluminosilicates, for the most part the potassic muscovite or the potassium-iron-magnesium biotite, both containing hydroxyl; the olivines, orthosilicates of iron and magnesium; nephelite, an orthosilicate of sodium and aluminum; leucite, a metasilicate of potassium and aluminum; magnetite, ferrosoferric oxide, often containing titanium; and apatite, a phosphate of calcium, containing a little fluorine or chlorine. The last two are present in almost all rocks but usually in very small amounts.

The average igneous rock.—Apparently F. W. Clarke[5,6] was the first to undertake to estimate the average chemical composition of the accessible earth's crust. Since then he and others, Harker, Mennell, Knopf, Mead and Washington, have published other estimates,* which do not differ greatly the one from the other.

The true estimation of the average chemical composition of the igneous rocks is by no means such a simple matter as it may appear to be at first thought. In the first place, we know but little of the exact chemical characters of the igneous rocks of many districts of the earth. This is true of the great continents of Asia and South America, as well as of Africa and Australia, concerning all of which we have, for the most part, a knowledge only of the rocks more or less near the coasts and know only in a general and very imperfect way the rocks that constitute the vast expanses of the interior portions. The same ignorance, either total or partial, holds true for many countries, such as China, Arabia and even Brazil, India, Egypt and Spain, for which the number of analyses is quite disproportionate to the number and masses of igneous rocks that are known to occur.

A second disturbing factor, and one that has been often advanced against the validity and representativeness of the estimates of the average composition of rocks, is that the true relative amounts of various rocks are not properly represented because of the selection of

* For literature citations see Ref. 3.

material for analysis. It has frequently happened that the petrographer has had analyzed rather the rare or most interesting rock types than those which, though much more abundant in the region described, are of more usual character. Although this is often to be expected and, from a special point of view, is almost justifiable, yet it certainly may involve a serious disturbance in the estimate of the composition of the crust as a whole. This is so because the most interesting types are usually less abundant than the common ones, so that in the determination of average composition they tend to be disproportionately represented.

TABLE 9

AVERAGE CHEMICAL COMPOSITION OF IGNEOUS ROCKS

(Weight per cent)

SiO_2	59.12	S	0.052
Al_2O_3	15.34	$(Ce,Y)_2O_3$	0.020
Fe_2O_3	3.08	Cr_2O_3	0.055
FeO	3.80	V_2O_3	0.029
MgO	3.49	MnO	0.124
CaO	5.08	NiO	0.025
Na_2O	3.84	BaO	0.055
K_2O	3.13	SrO	0.022
H_2O	1.15	Li_2O	0.007
CO_2	0.102	Cu	0.010
TiO_2	1.050	Zn	0.004
ZrO_2	0.039	Pb	0.002
P_2O_5	0.299		100.00
Cl	0.048		
F	0.030		

Although these objections are serious, they are not overwhelming. There are several factors, which cannot be discussed here, tending to compensate incorrect trends and to increase the probability that a general average of a large number of analyses will represent the composition of the outer crust.

The basis for the latest estimate is the comprehensive collection of 8,602 rock analyses compiled and published by Washington.[7] From 5,159 of these analyses Clarke and Washington[8] obtained the average composition shown in Table 9. From this table it will be seen that silica is the most abundant oxide in the crust, with alumina a poor second. The first nine oxides (from silica to water) constitute 98 per cent of the whole.

In terms of the minerals present the "average" rock found at the surface would have the approximate composition shown in Table 10. The crust, as a whole, corresponds to a granite or granodiorite.

TABLE 10

Composition of Average Igneous Rock in Terms of Minerals
(Weight per cent)

Quartz	11
Andesine	47
Orthoclase	16
Hornblende, Biotite	20
Magnetite	5
Apatite	1
	100

TABLE 11

Average Composition of Igneous Rocks
(Major and common elements)
(Weight per cent)

Oxygen	46.59
Silicon	27.72
Aluminum	8.13
Iron	5.01
Calcium	3.63
Sodium	2.85
Potassium	2.60
Magnesium	2.09
Titanium	0.63
Phosphorus	0.13
Hydrogen	0.13
Manganese	0.10
Sulphur	0.052
Barium	0.050
Chlorine	0.048
Chromium	0.037
Carbon	0.032
Fluorine	0.030
Zirconium	0.026
Nickel	0.020
Strontium	0.019
Vanadium	0.017
Cerium, yttrium	0.015
Copper	0.010
Uranium	0.008
Tungsten	0.005
Lithium	0.004
Zinc	0.004
Columbium, tantalum	0.003
Hafnium	0.003
Thorium	0.002
Lead	0.002
Cobalt	0.001
Boron	0.001
Beryllium	0.001
	100.00

TABLE 12

AVERAGE COMPOSITION OF IGNEOUS ROCKS

(Less common and rare elements)

(Relative proportion)

Cerium, yttrium	1.5×10^{-4}
Copper	1×10^{-4}
Uranium	8×10^{-5}
Tungsten	5×10^{-5}
Lithium	4×10^{-5}
Zinc	4×10^{-5}
Columbium, tantalum	3×10^{-5}
Hafnium	3×10^{-5}
Thorium	2×10^{-5}
Lead	2×10^{-5}
Cobalt	1×10^{-5}
Boron	1×10^{-5}
Glucinum	1×10^{-5}
Molybdenum	$n \times 10^{-6}$
Rubidium	$n \times 10^{-6}$
Arsenic	$n \times 10^{-6}$
Tin	$n \times 10^{-6}$
Bromine	$n \times 10^{-6}$
Cesium	$n \times 10^{-7}$
Scandium	$n \times 10^{-7}$
Antimony	$n \times 10^{-7}$
Cadmium	$n \times 10^{-7}$
Mercury	$n \times 10^{-7}$
Iodine	$n \times 10^{-7}$
Bismuth	$n \times 10^{-8}$
Silver	$n \times 10^{-8}$
Selenium	$n \times 10^{-8}$
Platinum	$n \times 10^{-9}$
Tellurium	$n \times 10^{-9}$
Gold	$n \times 10^{-9}$
Iridium	$n \times 10^{-10}$
Osmium	$n \times 10^{-10}$
Indium	$n \times 10^{-11}$
Gallium	$n \times 10^{-11}$
Thallium	$n \times 10^{-11}$
Rhodium	$n \times 10^{-11}$
Palladium	$n \times 10^{-11}$
Ruthenium	$n \times 10^{-11}$
Germanium	$n \times 10^{-11}$
Radium	$n \times 10^{-12}$

Tables 11 and 12 show the average amounts of the *chemical elements* in igneous rocks. Of the known 92 elements, 75 are listed here. It is seen that the most abundant element in the crust of the earth, oxygen, makes up rather less than one-half (46.59 per cent) of the crust; silicon, the second most common element, accounts for a trifle over one-

quarter, with aluminum, third, comprising about one-twelfth and iron, fourth, about one-twentieth. Calcium, sodium, potassium and magnesium follow in this order, with no great differences among them (3.63 to 2.09 per cent). Then comes titanium, the ninth on the list, with a percentage of about 0.63 per cent. These 9 elements, then, make up about 99.25 per cent of the earth's crust. After them come small amounts of phosphorus, hydrogen and manganese, these 12 elements constituting 99.61 per cent, leaving only about 0.39 per cent for all the other elements.

Some very striking facts are shown by these tables. One is the very small number of elements that make up more than 99.50 per cent of the crust of the earth. Another is the absence from the list of the most abundant elements in the earth's crust of so many elements and metals that are necessary to our daily need and essential to our civilization. Thus lead, tin, copper, zinc, mercury, gold, silver, platinum, antimony and arsenic are not represented. All these are found in igneous rocks, but only in scarcely detectable amounts, and they are made available to our use only by processes of natural concentration into so-called *ore bodies.* Of the metals in common use only iron, aluminum and manganese are on the list of the first 12.

Another striking fact, first brought out by W. F. Hillebrand, is the relative abundance of titanium. This element was, for many years, considered to be rare and was not determined in making a chemical analysis of a rock. But, nowadays, no rock analysis is considered as being satisfactorily complete if it is not determined.

VOLCANIC ACTIVITY

The phenomenon of volcanism[9] has an important bearing on the nature and state of the earth's crust and of its interior, and furnishes direct and striking evidence of the conditions many miles below the surface. The existence of volcanoes and the efflux of vast flows of lava from them or from fissures were formerly, and still are popularly, brought up as an argument against the notion of the solidity of the interior. The present-day view, however, is that these are, to a large extent, local phenomena, of shallow origin, and that the existence of such deeply buried masses of liquid magma as these may be does not affect the essential rigidity of the earth.

Almost all substances, with the exceptions of water, bismuth and a few others, contract on solidification from the liquid state or, conversely, expand on melting. The melting point, therefore, in accordance with Le Châtelier's law, is raised by pressure. Many

investigations,[10–22] more especially in recent times
Joly, Douglas, and Day, Sosman and Hostetter, have shown t
igneous rocks and the minerals of which they are composed belong to this category; *i.e.*, they contract on solidification, so that under sufficient pressure they will be solid at temperatures at which they would be liquid under atmospheric pressure (see section on Crystallinity). It follows, therefore, that at a certain depth below the surface, dependent on the composition of the magma, its gas content and other factors, a mass of magma, although kept solid by pressure, may be potentially liquid, so that, on release or diminution of pressure, it may liquefy (melt) and flow as a lava and thus reach the surface of the earth or form a laccolith below the surface.

The modern concept of the sources of volcanic lavas and of batholithic intrusions is based on this supposition. It is frequently stated that lavas originate from more or less isolated masses (so-called *reservoirs*, *pockets*, *basins*, *maculae*, etc.). But these are essentially the result of a localized diminution of pressure, brought about by crustal movements, from which the actually solid, but potentially fluid, magma may flow and escape at the surface as lava if a way is open. As Iddings expresses it:* "A primitive 'magma basin,' then, is not a vessel of liquid, but a condition of the Earth, obtaining possibly at various unknown distances below the surface, according to circumstances of composition, temperature, and pressure."

Many observations on volcanic eruptions and on volcanic quakes indicate that these feeding masses of magma, the loci of diminished pressure, which are the uppermost expression of deeper lying material, are situated not far below the surface, at depths of the order of 10 km. or less. They are so isolated, sporadic and discontinuous (although possibly connected with much more extensive regions of magma below), at such shallow depths and of such relatively small volume, that they do not affect the rigidity of the earth as a whole. To human beings, volcanoes (and their feeding basins) may appear to be immense: considered as part of the whole earth they may be quite insignificant, surficial phenomena.

TEMPERATURE OF LAVAS

The temperature of lavas as they issue at the surface is of interest and importance. Up to the present time, observations have been made only on basaltic lavas; the temperature of andesitic and rhyolitic lavas is as yet unknown, although it is reasonable to suppose that it

* Ref. 19, p. 464.

does not differ much from that of basalts, as the greater viscosity of the more silicic lavas is offset by the presence of dissolved gases.

The temperature of the lava lake at Kilauea has been measured by thermocouple and optical pyrometers and by Seger cones,[23–26] with concordant results; it is found to be as high as about 1200°C., with an average of about 1100° a few meters below the surface. The temperature of flows of leucite tephrite at Vesuvius[27,28] has been found to be about 1100°, and similar temperatures have been observed at Etna.[29,30] It has also been estimated that the intrusion temperature[31] of a New Jersey diabase was not above 1150°. A temperature of about 1100° for basaltic lavas when they reach the surface will, thus, be probably not far wrong.

It should be noted, however, that, although part of the heat of molten lava is undoubtedly intratelluric, part of it may be due to two other causes, *viz.*, exothermic interreactions among the dissolved gases which take place with diminishing pressure,[32–36] and effects related to crystallization.[37]

Volcanic emanations afford perhaps the most direct evidence of elevated temperatures in the interior, and this conclusion is confirmed by the existence of hot springs and by the increase in temperature in deep mines, tunnels and bore holes. The temperature increases continuously as the depth increases down to about 4 km., the limit of our observed knowledge, beyond which depth the temperature is quite unknown.[38–44] The temperature gradient differs much in different places even for this slight depth, which is about $\frac{1}{3000}$ of the earth's radius, and in the same bore hole it may vary with depth. The average gradient is variously stated, usually as about 1°C. for every 30 m. (about 1°F. for every 60 ft.), but it may be much more or much less than this (see Chap. VI). Daly[45–46] has pointed out that the average temperature gradient for European bore holes appears to be less than the average for North American bore holes. The average of the data tabulated by Van Orstrand,[47] however, indicates that the American and European gradients are nearly equal.

The gradient may be much influenced locally by several factors, *e.g.*: the kind of rock and hence its heat conductivity; the proximity of hot springs or masses of still hot, intruded igneous rock; the effect of subterranean streams and the circulation of water; the wetness or dryness of the rock; the presence of ore bodies and rise in temperature through the oxidation of sulphides, etc.; the presence of radioactive elements; the effect of increasing pressure and density in changing the heat conductivity. A more detailed discussion of temperature gradients will be found in Chap. VI.

CRYSTALLINITY

The material of the continental layers is almost entirely holocrystalline.* This is shown by the holocrystalline texture of the plutonic rocks and of many of the effusive lavas, the relative amount of glassy lavas at the surface being insignificant as compared with volume of the whole crust. For the portion below the visible crust, the same condition of holocrystallinity and heterogeneous mineral composition is in accord with reasonable estimates of the temperatures at the bottom of the crust, say at 60 km. If the observed average temperature gradient of about 1°C. for each 30 m. continues unchanged down to 60 km., the temperature at this depth would be about 2000°. It is, however, probable that the gradient decreases with depth, so that the temperature at 60 km. is considerably less than 2000°; and furthermore the pressure at this depth, about 17,000 atmospheres, might easily raise the melting point of the material to such a degree that there is no difficulty in imagining the material to be solid. On the whole, although there is no definite proof, it seems probable that the crust is holocrystalline down to a considerable depth.

With respect to the condition of the deeper parts, there is much more uncertainty. Several important lines of reasoning indicate that the material in this region is largely or wholly glassy. The tremendous volume of the great basaltic flows has appeared to require a deep-seated reservoir of molten or glassy material of almost limitless extent.

BASALTIC LAYER

The continuous increase in temperature with increase in depth is held to indicate, with moderate and reasonable extrapolation, that at very moderate depths, say at 100 km. or so (and a fortiori below this), the material must be at a temperature above its melting point, although in a state of rigid solidity or high viscosity because of the pressure to which it is subjected. It is natural to conceive of matter in such a condition to be glassy and noncrystalline, as pointed out by Daly,[46] although a crystalline state is not necessarily excluded. The conditions in this region are so extreme and so far removed from those attainable by our present laboratory facilities that nothing very definite can be said.

* The term *crystallinity* is used in the sense of *degree of crystallization* of a rock, as contrasted with a *state of glassiness*. See W. Cross, J. P. Iddings, L. V. Pirsson, H. S. Washington. Jour. Geology, **10**: 611 (1902); and "Quantitative Classification of Igneous Rocks." P. 154. Chicago, 1903.

It would seem, however, that the very extensive and profound movements in the crust, shown by faults, thrusts, warping, folding and mountain building, as well as by the condition of isostatic adjustment, "necessitate that, below the rigid and solid crust, must come material which possesses some of the properties attributed to a fluid, though not necessarily more than the power of changing its form when exposed to stresses of sufficient magnitude and of long enough duration"[48] (see Chap. XV). To this zone, Barrell[49] has given the name asthenosphere—the zone of weakness. This is essentially analogous to, although more viscous than, the liquid *undercrust layer*, suggested many years ago by Dana[50] and others, to account for orogenic and other crustal movements.

The idea that a basaltic substratum underlies the granitic crust was first suggested (in 1844) by Charles Darwin,[51] who, arguing from the accumulation of the heavier crystals at the bottom of lava flows and from the abundance of dikes of greenstone and basalt in many regions of granite, concluded that trachytic rocks overlie basaltic. Von Cotta later[52] developed in greater detail the suggestion that a basaltic substratum underlies the upper granitic layer; since his time this concept of the crust has had many advocates, one of the most active being Daly,[53] who has done much to advance the idea, which is now generally accepted.

This vertical distribution or differentiation is supposed to have been brought about in great part by the sinking of the heavier minerals, or the rising of the lighter, in the originally liquid outer layer of the then molten globe, when the surface began to solidify. This idea is expressed in Daly's terms *gravitative adjustment* and *gravitative differentiation.* Laboratory experiments that demonstrated this action were made by De Drée prior to 1825,* and the more extensive modern investigation by Bowen[54] have shown it conclusively. Some investigators have held that there is a danger of exaggerating the importance of this action and that there is a possibility of other processes being involved.

The composition of the material in the presumable deep-seated basaltic gabbroic layer may best be inferred from the composition of plateau basalt. The latest average given by Daly† for "world plateau basalt" is probably the most useful and is reproduced in Table 13. The corresponding theoretical mineral composition is as follows: plagioclase

* See Von Buch, Leopold, Physikalische Beschreibung der Canarischen Inseln. P. 230. 1825. Also, Darwin, Charles, Geological Observations. 3d ed. P. 133. 1891.

† Ref. 46, pp. 17, 201.

(albite and anorthite in the proportion 3 to 2), 50; augite, 37; olivine, 4; and "ores," 9.

The depth of the assumed basaltic layer has been estimated in various ways. Seismologists, from the study of near-by earthquakes, have inferred a decided change in the character of the material at depths of 30 to 50 km. The evidence on which these estimates rest is of different kinds: the depth at which reflection of the two waves takes place, the change in velocity of the waves, etc.

TABLE 13

AVERAGE CHEMICAL COMPOSITION OF PLATEAU BASALT

(Daly)

SiO_2	50.60
Al_2O_3	17.40
Fe_2O_3	4.57
FeO	6.29
MgO	4.89
CaO	8.09
Na_2O	3.23
K_2O	1.76
H_2O	1.83
TiO_2	0.68
P_2O_5	0.20
MnO	0.46
	100.00

The depth of the focus, or place of origin, of earthquakes as well as volcanic eruptions was formerly thought to be not far below the surface,[55–59] say 30 km., but earthquakes are now known to originate at depths of several hundred kilometers (see Chap. XI). That earthquakes are caused by movement along fracture surfaces implies a solid capable of sudden or almost instantaneous rupture. The existence of comagmatic regions, the persistence during the life of a volcano of a definite, unchanging kind of lava (Mt. Pelée, Santorini) or the orderly succession of closely related kinds (Vesuvius, Etna, Pantelleria, Monte Ferru, Mauna Loa) and other such features indicate that the magma basins are isolated and hence imply that they are enclosed by solid rock and not directly connected with a terrestrially large and general source. It seems difficult for the depth of focus of earthquakes and of volcanic eruptions to yield much information concerning the depth of the major layers in the crust.

From study of gravity anomalies and isostasy in the United States, Hayford and Bowie[60–65] have variously estimated the *depth of isostatic compensation* to be 122 to 60 km. below the surface. The value adopted by Bowie is 96 km., based on determinations of gravity in

mountainous regions of the United States, whereas the value 60 km. is derived by him from gravity determinations made at localities all over the United States.

A study of the relations of the average density of the igneous rocks of continents, ocean floors and smaller areas over the globe to their mean altitude led Washington[66] to the estimate of 50 km. as the depth of the *isopiestic level, i.e.*, that of equal pressure, a value which is practically identical with Bowie's more general estimate. Daly* arrives at an estimate of about 40 km. for the thickness of the crust or perisphere, basing his conclusion on several lines of evidence, thermal gradients, seismology, cosmogony and geodesy. Goldschmidt[67] thinks that the diamonds in the kimberlite pipes of South Africa, which he supposes to have been derived from an underlying eclogitic layer, "must have been formed at a depth of at least 60 kilometers."

Other estimates have been made, based on various other lines of evidence, but they are all of the same order of magnitude. The thickness of the perisphere is almost certainly not uniform and is presumably indefinite. Just as its upper surface shows great variation in altitude above sea level, so the under surface cannot form a uniform spherical surface. The individual thicknesses of the presumably granitic and basaltic layers will be treated in Chap. XII.

It is not known whether the crust grades continuously into the underlying mantle or whether they are separated by a discontinuity. Several considerations call for continuity at this junction; others, based on the study of earthquake waves, imply either a discontinuity or a change in physical state within a very short distance. The trend of evidence indicates that the last hypothesis is nearest to the truth, *i.e.*, that, while there is no absolute discontinuity between crust and mantle, the great change in physical (and chemical) characters takes place within a relatively slight distance.

It is supposed by some that the upper shell of the "interior" is much more "basic" in chemical composition, *i.e.*, peridotitic. By others this zone is thought to be not very different in composition from a typical basalt. This would be Daly's (Ref. 18, p. 171) "basaltic substratum which is still hot enough to flow" or his "homogeneous vitreous shell underlying a heterogeneous crust." It is upon such a shell of highly viscous glass that Taylor,[69] Wegener,[70] Daly and others imagine the North and South American continents to have slid westward from Europe and Africa—an idea that has provoked much adverse as well as favorable criticism[70–76] (see Chap. IX).

* Ref. 45, p. 371.

In any case, in order to account for the possibility of the extensive and profound faulting, dislocation, warping, thrusting, diastrophism and other movements that the crust has undergone, it seems to be necessary to assume a mantle that is capable of yielding to sufficiently great and sufficiently long-continued stresses and that is, at the same time, seismically isotropic and sufficiently rigid to be capable of transmitting both longitudinal and transverse earthquake waves, as well as of meeting the demands for rigidity already mentioned.

DENSITY

The density of the accessible crust (epiperisphere), assumed to be composed wholly of known igneous rocks in the proportions in which they are found at the surface,[7] is about 2.79,[8] *i.e.*, about one-half of the earth's density. The value would be very slightly less than 2.79, possibly 2.77, if a small amount of sedimentary rocks is taken into consideration. Earlier estimates are generally lower than 2.79, Fisher[1] accepting the value 2.75 whereas American geodesists adopt Harkness' value 2.67. The mean density of the earth as a whole is about 5.52[77–79] (see Chap. XIII).

It is generally* assumed that the density of the earth increases regularly with increase in depth from the surface to the center, this increase being brought about partly by the continuously increasing pressure and partly by change in composition. But there is considerable divergence of opinion as to the rate of increase, as well as whether the increase is continuous or discontinuous. Consequently there is diversity in the various estimates that have been made of the density of the matter at the center of the earth. The early estimates of the density of the central matter vary from the density of the surface rocks to infinity.[81] Modern estimates are more consistent. Both Clarke[82] and Farrington[83] assume the density 7.8 for that of the central core of nickel iron; Goldschmidt[84] assumes a density of about 8; according to Wiechert[85] the density of the core is 8.4; whereas Williamson and Adams,[80] who take into account the effect of compression and other factors, assign an average density of about 10 to the central nickel-iron core, which would (through compression) attain a value of about 10.7 at the center (see Chap. XIII).

MAGNETISM

The earth as a whole acts in many respects as a magnet. Inasmuch as the rocks of the crust in general are not notably magnetic, this

* Hobbs imagines an innermost core of ironstone of density 6.93, surrounded by a zone of nickel iron of density 7.6: Daly assumes a stratum of basaltic glass, of density 2.75, underlying a suboceanic zone of basalt of density 2.95.

action may be attributed, in great part at least, to the magnetic character of the material that forms the interior, although part of the effect may be due to electric currents in the earth or to its rotation.[86–89] It is natural to suppose that the internal magnetic material is iron or nickel iron, especially as other lines of evidence, such as analogy with the meteorites, lead us to believe in the existence of a central core or kentrosphere of nickel iron.

Bauer[90] states that 94 per cent of the earth's total magnetism is due to causes beneath the surface, the remaining 6 per cent originating in the atmosphere or in space about the earth. He also found by comparing values obtained from his analysis for the epoch 1922 with those for earlier epochs that there had been an apparent decrease in the average equivalent magnetization of the earth for the preceding 80 years at an average annual rate of 1/1,500 part. Of course, this apparent decrease may arise from inaccuracies in the earlier data. According to Bauer: "If the Earth's magnetism were distributed uniformly throughout its volume, as it probably is not, the average intensity of magnetization would be 0.074 C.G.S. The magnetic axis intersects the north hemisphere in latitude 78° 32′N. and longitude 69° 08′W. of Greenwich."

A very interesting relation brought out by Bauer's studies is that the magnetic intensity is apparently dependent on, or connected with, the distribution of the great land and water areas of the globe, in the sense that "the average intensity of magnetization for corresponding parallels north and south is generally larger for the land-predominating parallel than for the ocean-predominating parallel." In other words, the internal magnetic field is not symmetrical about the geomagnetic equator, so that it is represented by a pear-shaped solid. The upper encircling bulge of this corresponds with the zone north of the equator, up to about lat. 30°N., which is the longitudinal zone of the maximum land masses. South of the equator, the encircling trough, from about lat. 20 to 60°S., corresponds to the longitudinal zone of the greatest oceans. Also, beneath the Arctic Ocean there is an axial depression, while there is a marked protuberance beneath the Antarctic continent.

METEORITES

Many different origins have been assigned to meteorites, such as that they have been thrown out from the sun or from volcanoes of the earth or of the moon, that they are portions of comets or fragments of the asteroids or are clouds of cosmic dust condensed in the atmosphere. Almost all these suggested origins have been abandoned, because of the serious objections that may be brought up against them. The view

now generally held, with slight modifications by some,[91–95] is well expressed in the words of Farrington:[96] "Meteorites are portions of a disrupted mass of cosmic matter, which had a spherical form, increased in density toward the center, and cooled from a liquid or semi-liquid to a solid state before disruption." The chief facts advanced by Farrington are enumerated below; it is needless to discuss them in detail here. The present writer cannot but agree with Clarke[82] in his opinion that "Farrington's arguments in support of his thesis seem to be incontestable." The bases of Farrington's arguments are as follows, slightly altered in form:

1. Most iron meteorites are octahedral in crystallization.
2. Most stone meteorites are chondritic and contain glass.
3. There is every gradation between iron and stone meteorites.
4. The substance of most meteorites was in a solid state before their fall upon the earth.
5. The structure of most meteorites shows that their substance has changed from a liquid to a solid condition.
6. The structure of most iron meteorites shows that the change from a liquid to a solid state has taken place slowly.
7. The structure of most stone meteorites shows that the change from a liquid to a solid state has taken place rapidly.
8. The high temperature (gamma) form of iron crystallizes octahedrally.
9. The carbonaceous meteorites are of exceptionally low specific gravity.
10. Diamonds are found in some meteorites, and may be produced artificially by crystallization, slowly and under great pressure, from carbon dissolved in molten iron.

Other features of meteorites that bear on an understanding of their origin are:

11. The presence of nonoxide compounds, such as carbides, phosphides and sulphides, in both iron and stone meteorites.
12. The abundance of metallic nickel iron.
13. The absence of minerals containing water or hydroxyl and (except very doubtfully) carbon dioxide.

Since the publication of Farrington's paper some features of meteorites have been observed, or those then known have been further studied and results obtained that lend additional support to Farrington's view as to their origin. Among these are examples of brecciation, slickensides, veins, tuffaceous structure and metamorphism;[97] and the interpretation of most chondrules as solidified drops of "fiery rain"[98–99]

further strengthens his theory. These and other characters imply that meteorites were formed under conditions possible only in a body of very considerable dimensions, probably much smaller than the moon but certainly vastly larger than even the largest of the single meteorites or the aggregate mass of any of the meteoritic showers that have reached the earth. It is not known whether meteorites are fragments of but one such body or (more probably) of more than one or what their relations are to the solar system:* discussion of these topics is not germane to the present paper.

The chief mineral constituents of meteorites are,[93,94,100] in the order of their abundance: nickel iron, olivine, rhombic pyroxene (enstatite, bronzite, hypersthene), monoclinic pyroxene (augite, diopside, hedenbergite, clinohypersthene), plagioclase (anorthite, labradorite, oligoclase) and maskelynite (probably fused feldspar). Schreibersite (iron-nickel phosphide), troilite (ferrous sulphide), chromite, cohenite (Fe_3C), graphite and tridymite are often present in small amounts. Diamond has been found in several irons, and some meteorites contain small quantities of amorphous carbon and hydrocarbons. Many irons contain lawrencite (ferrous chloride, $FeCl_2$) and small amounts of nonterrestrial minerals, oldhamite (CaS), daubreelite ($FeS \cdot Cr_2S_3$), moissanite (SiC); asmanite, a form of tridymite (SiO_2), also rarely occurs.

There are some interesting and significant features of the mineralogy of meteorites as compared with terrestrial minerals. Nickel iron, the most abundant meteoritic mineral, is very rare terrestrially; but olivine, the pyroxenes, both orthorhombic and monoclinic, and plagioclase are common in the rocks of the earth. On the other hand, such very common terrestrial minerals as orthoclase, nephelite, leucite, the micas and amphiboles, garnet, tourmaline, scapolite and a host of others are never found in meteorites; and quartz, the low-temperature form of silica, the most abundant earthly mineral, is an exceptional rarity in meteorites.[101] In general, the aluminosilicates and minerals that contain water, hydroxyl or (probably) carbon dioxide are not found in meteorites. In brief, the most abundant minerals of meteorites, except nickel iron, are olivine, pyroxene and calcic plagioclase, which compose terrestrial "basic" or "ultrabasic" (dofemic or perfemic) igneous rocks, such as peridotite, dunite, pyroxenite and very femic gabbro. The minerals that largely make up granite, syenite, diorite and their effusive equivalents—rocks that are especially abundant in and characteristic of the outermost crust—are not found in meteorites. Also, the minerals that are most

* Berwerth suggests that they are derived from the asteroids.

characteristic of terrestrial pegmatitic, pneumatolytic and metamorphic processes and of weathering (prior to reaching the earth) are not present in meteorites. These features, so far as their aggregate chemistry is concerned, are expressed concisely in the average chemical composition of stone meteorites, which has been calculated by Merrill[102] and by Farrington.[103] The averages closely resemble each other, and, if the presence of metallic nickel iron is disregarded, they are very like analyses of terrestrial peridotitic rocks.

The characteristic occurrence in meteorites of nonoxide compounds, which are not known to occur on the earth, is another significant feature. It has been pointed out by many students of meteorites that such compounds could have been formed only in the absence of oxygen and water or under nonoxidizing conditions as otherwise they would have been changed to sulphates, phosphates,[8,82] etc. The presence of nickel iron, graphite, diamond, amorphous carbon and lawrencite (ferrous chloride), all found terrestrially, also points in the same direction.

The distribution of some of the less abundant minerals between iron and stone meteorites is of interest. Thus schreibersite [$(Fe,Ni)_3P$], cohenite (Fe_3C), graphite (C) and, to a less extent, lawrencite (F_2Cl_2) and moissanite (SiC) are most abundant in the irons; amorphous carbon (C), oldhamite (CaS) and chromite ($FeCr_2O_4$) are most abundant in stones; troilite (FeS) is about equally common to both. Schreibersite is, par excellence, the most characteristic and almost never-failing accessory mineral of the irons, as chromite is of the stones, whereas troilite is the most abundant accessory in all kinds of meteorites.[93,100] The gases that are occluded in meteorites also show discrimination, carbon dioxide being most abundant in the stones and hydrogen and less carbon monoxide in the irons.

Prior[104] has pointed out that "for any meteoric stone the richer in nickel is the nickel-iron, the richer in ferrous oxide are the magnesium silicates, or, in other words, the ratio of magnesia to ferrous oxide in the magnesium silicates varies directly with the ratio of iron to nickel in the nickel-iron." Thus, if the dispersed nickel-iron phase is low in nickel, the pyroxene and olivine are high in magnesia, and if the metal is high in nickel the silicates are high in iron. This view is supported by many analyses. Prior's explanation is that iron is much more readily oxidizable and "silicatizable" than is nickel, so that, in successive stages of oxidation after the primary formation of the metallic alloy, there is an increasing oxidation of the iron and, hence, an increasing formation of iron silicate, "since little or no oxidation of nickel would occur so long as any iron remained unoxidized."

Another significant fact in the chemicomineralogy of meteorites is that, whereas the metasilicate pyroxene group (mostly orthorhombic) and the orthosilicate olivine group are present in almost equal amount in stone meteorites generally,* olivine is far more abundant than pyroxene in the siderolites, the meteorites that contain about equal amounts of metal and silicate. On the other hand, pyroxene is more abundant than olivine in the achondritic stones, which contain little or no nickel iron and most resemble terrestrial ultrabasic rocks. This is especially evident in those that are monomineralic, about half a dozen being known that are composed wholly of orthorhombic pyroxene and only one that is composed of olivine alone, with a little chromite.

A most important feature of meteorites is that all known meteorites are of igneous origin; none is known that even remotely approaches in composition or texture a terrestrial sedimentary rock, such as sandstone or shale, or that has the texture of a terrestrial metamorphic rock, such as gneiss or schist. A small number of known stone meteorites have gabbroic texture, and these in many respects closely resemble terrestrial plutonic rocks of similar mineral and chemical composition. A much greater number are clastic, being igneous tuffs and breccias, which differ from terrestrial volcanic clastic rocks chiefly in the presence of sporadic nickel iron and of the peculiar, small, spheroidal, silicate bodies called *chondrules*. The iron meteorites (siderites) and most of the types of iron-stone meteorites (siderolites), such as pallasite, are not known to be represented terrestrially; but they are all unquestionably of igneous origin.

Classification.—In the classification of meteorites, the two chief factors, especially as regards their relations to the constitution of the interior of the earth, are the relative amounts of nickel iron and silicates and whether the metal is sporadic (dispersed) in the silicate or the silicate is sporadic in the metal.[105] Meteorites may thus be classified broadly as follows,[93,94,103,104,106–109] a classification that is in accordance with modern practice, except that emphasis is laid, for our special purpose, on whether the iron or the silicate is sporadic, *i.e.*, in scattered, discrete particles.

1. *Aerolites*, stony meteorites, which are composed almost wholly of silicates, mostly olivine and pyroxenes. There are two textural divisions. (*a*) The chondrites contain chondrules, which are small, rounded bodies, mostly of silicates and found only in meteorites. The chondrites are composed mostly of olivine and pyroxene, many of them

* Tschermak found the average ratio of olivine to bronzite to be 9 to 8 in 66 meteorites (Farrington, Ref. 93, p. 182).

with glass; most of them have a tuffaceous texture, and nearly all of them contain nickel iron, up to about 25 per cent, sporadic in small particles throughout the continuum of silicates. A few chondrites are carbonaceous, containing amorphous carbon and hydrocarbons, although the stone is evidently of igneous origin. (*b*) The achondrites are the second kind of aerolite. These are without chondrules and are composed in large part of pyroxene or olivine or both, with little or no nickel iron. The achondrites have generally a crystalline-granular texture, with little or no glass; they are usually not tuffaceous, and most of them closely resemble some basic or femic terrestrial plutonic igneous rocks.

2. *Siderolites*, stone-iron meteorites, consisting of nickel iron and silicates, both in large or approximately equal amount.* In almost all siderolites, the silicate forms rather large, rounded grains sporadic in a continuous sponge of nickel iron; in some siderolites the metallic sponge is more or less discontinuous, but few siderolites are known in which the metal is definitely sporadic in the silicate. In the greater number of siderolites the silicate is olivine, mostly without pyroxene, this combination of metal and olivine forming the rather abundant pallasite; in fewer siderolites both olivine and bronzite are present; in still fewer is the silicate entirely pyroxenic. Chondrules are rare.

3. *Siderites*, iron meteorites, composed almost wholly of nickel iron, nearly always with schreibersite and less often with troilite and graphite, these accessory minerals forming small, rounded masses sporadic in the metal base. The structure of the metal in most siderites is highly crystalline, in the greater number of octahedral, brought out on etching (Widmanstaetten figures), in fewer cubic and in still fewer not evident. The nickel iron is of three kinds, which differ in their mode of crystallization, their content in nickel and in other ways.

4. *Tektites.*†—Besides the meteorites that are universally recognized as such, there are some rare and peculiar glassy bodies of which the meteoritic origin is probable but is still in dispute.[109a–118] These are composed entirely of glass, without inclusions or crystals, always small, sometimes irregular but more often rounded and curiously pitted, in color black, green, yellowish or colorless. They are found, scattered on or near the surface, within only a few restricted areas, in

* In a recent study of 17 pallasites, Chirvinsky determined the average percentage of olivine to be 51, ranging from 37.18 to 75.11; see the abstract in Min. Mag., **20**: 83 (1923).

† The tektite has been given several other names: moldavite, bilitonite, australite, obsidianite, etc.

Bohemia, Moldavia, the Dutch East Indies, the Malay States, Australia and Tasmania; none seems to have been found in the western hemisphere* or in Europe or Asia far from the narrow belt along which the localities lie.

Various origins have been attributed to the tektites: that they are artificial products, volcanic *ejectamenta*, possibly volcanic bubbles, rolled obsidian pebbles, products of the fusion of atmospheric dust by lightning, and that they are meteoritic. Of these, the meteoritic origin is the best founded and most probable—indeed a meteoritic origin may be safely accepted for them. The main reasons for this view are well stated by Summers in his paper of 1913; their occurrence in few and relatively small areas, scattered much like showers of meteorites; their occurrence only on or near the surface; the absence of recently active volcanoes in their vicinity, especially any furnishing lavas of similar composition; their peculiar forms and surface markings; and their unusual and rather uniform chemical composition, which is unlike that of terrestrial obsidian.† This peculiarity in chemical composition consists in the conjunction of high silica with high alumina, potash and lime and low magnesia, iron oxides and soda. The study by Tilley of the refractive indexes and densities of the tektites also indicates clearly their marked dissimilarity to terrestrial volcanic glasses. The argument from surface markings is inconclusive of either a meteoritic or a terrestrial origin, as they might be produced on fallen tektites, just as they are on terrestrial obsidian pebbles, by natural etching[112,119] [see Merrill, U. S. Nat. Mus. Proc., **40**: 481 (1911)].

It is therefore reasonable to think that the tektites, like the true meteorites, are fragments of cosmic bodies of which they presumably formed part of the thin outermost crust. Their comparative rarity and small size would be consonant with this origin, as such a thin crust would be of small volume relative to the mass of the planetoid and the fragments of glassy material would be readily disintegrated during the passage through the atmosphere.

The chemical composition of the tektites, in which they resemble a few peculiar terrestrial granites, and their constant wholly vitreous texture, as well as the absence of meteoric sedimentary rocks, lead to the speculative suggestion that the cosmic bodies from which they were derived had, like the moon, no atmosphere because of their small mass. Consequently the surface was not subjected to weathering or

* G. Linck (Aufbau des Erdballs. P. 11. Jena, 1934) reports a tektite from Peru.

† For analyses of tektites see Ref. 7, pp. 51, 53, 55, 103, 107, 331 and 333.

erosion, and the tektites represent the original glassy skin formed at the time of their solidification, such a glassy skin as the earth must have had at an early period, but of which all traces have been lost through erosion brought about by the presence of an atmosphere and erosion by water on the surface.

An interesting and very suggestive feature of the localities in which tektites are found is that all the small scattered areas, from Bohemia to Tasmania, are along an arc of a great circle, the length of which is about 150°, cutting the ecliptic in long. 110°E. at an angle of about 40°. This suggests the possibility that the earth picked them up during its passage centrally through a thin stream or sheet of these bodies. The total arcual length, on this supposition, should be 180°, but the lesser actual length is readily accounted for by the presence of the expanse of ocean southeast of Tasmania. The Bohemian localities would thus seem to be at the northern extremity of the semicircular arc, as no tektites have been found in the densely populated country to the northwest. That the localities are grouped in three distinct and widely separated regions indicates that the matter composing the stream or sheet was more or less bunched. If this idea as to a stream or sheet of tektites is true, the falls at the different localities must have been contemporaneous or, rather, nearly simultaneous.

Average composition of meteorites.—In Table 14 are shown the averages for various types of meteorites in weight per cent. The average composition of 318 iron meteorites and that of 125 stone meteorites were obtained by Farrington.[103] For comparison, Merrill's average[102] for 53 stone meteorites is also given. The agreement with Farrington's results is excellent. In the last column are shown the average composition of 20 achondritic meteorites, which was obtained by Washington.[98] It has been suggested by Farrington[103] that the composition of the earth as a whole is the same as the average for all meteorites. For the purpose of obtaining a proper average it is preferable to take the iron and stony meteorites, not in the proportion in which they are found, but in the proportion in which they are seen to fall. The ratio is approximately 35 stony meteorites to 1 iron meteorite and is so large that the general average on this basis would not differ very much from the average for the stony meteorites. In terms of minerals this corresponds to: olivine, 35; pyroxene, 42; anorthite, 4; troilite, 5; nickel iron, 13. The silicate portion is principally an olivine-pyroxene mixture, *i.e.*, essentially a peridotite.

Native iron.—Flows and sheets of basalt and diabase at several localities contain metallic iron or nickel iron. The most important of these localities are: Disko Island and elsewhere on the west coast of

TABLE 14
COMPOSITION OF METEORITES
(General average, 35 stone to 1 iron)

	Iron, average of 318	Stone, average of 125	Stone, average of 53	Achondritic, average of 20
1				
Fe (metallic)	90.67	11.46	12.18	1.18
Ni (metallic)	8.50	1.31*	1.57*	0.33
Co (metallic)	0.59	0.05	0.07	0.04
O	......	36.02	37.10	42.05
Si	......	18.41	18.34	23.00
Al	......	1.39	1.55	3.26
Fe (silicate)	......	12.88	10.37	12.33
Mg	......	13.54	13.88	10.91
Ca	......	1.65	1.72	5.09
Na	......	0.59	0.65	0.50
K	......	0.17	0.14	0.22
Ti	......	0.01		
Cr	......	0.28	0.34	0.31
Mn	......	0.14	......	0.18
S	0.04	1.98	1.82	0.54
P	0.17	0.06	0.12	0.06
C	0.03	0.06	0.15	
Total	100.00	100.00	100.00	100.00

* Includes a small amount of Ni as NiO.

Greenland;[120–124] the Buhl, near Kassel, Hesse-Nassau, Germany;[125] the small, recent, extinct volcanoes near Olot, in Catalonia;[126,127] the "traps" of New Hampshire;[128] Volhynia, Russia;[129] the district of Okhotsk in eastern Siberia.*

The iron in all these occurrences is wholly sporadic, much as it is in the chondrites,[105] nothing approaching the pallasite texture having been noted. There is great variation in the amount of the metallic iron, as well as in the size of grain. At Disko, masses of iron occur, up to 25 tons in weight, and at Buhl up to several kilograms, at both localities with very much smaller grains, down to 1 mm. or so across. In general, as in Catalonia, in New Hampshire, and in Volhynia, the iron is in very small grains, to be detected only in thin section by the use of copper sulphate solution. Some of these native basaltic irons contain nickel, always less than in meteorites, that of Disko up to about 3 per cent, with a little cobalt; that of Buhl contains

* Oral communication by Dr. J. J. Sederholm.

neither nickel nor cobalt, according to Irmer, but further examination would be desirable. The iron of Ovifak contains carbon. Widmanstaetten figures, which are so characteristic of meteoric iron, are rare but are seen in some of the larger masses, especially in those of western Greenland.

The origin of this native iron has been attributed by many to reduction of magnetite or ferrous silicate by coal beds traversed by the flows. But this explanation is not altogether satisfactory, because of the presence of nickel and cobalt, the traces of crystalline structure and for other reasons; so that a truly primary occurrence of native iron in the basalt magma may be accepted for some of them at least.

Besides the basaltic iron, which is poor in nickel, there occur several native nickel-iron alloys that are very rich in nickel. Such are awaruite found in New Zealand, josephinite from Oregon, and souesite from British Columbia. The composition of these undoubtedly terrestrial and primary nickel irons, which are associated with serpentine (the decomposition product of dunite or peridotite), is rather variable but approximates to $FeNi_2$ or $FeNi_3$.

Abundance of iron in the crust.—Recent calculations of the average chemical composition of the igneous rocks that constitute the known outer crust[130–132] agree in showing that, expressed in terms of elements, iron is the fourth most abundant element in the igneous rocks of the crust, its relative amount being about 5 per cent; only oxygen, silicon and aluminum are more abundant. As there is good reason to believe that rocks richer in iron, such as gabbro, basalt and peridotite, are more abundant not very far below the prevailingly granitic rocks of the uppermost 10 miles, the abundance of iron at the surface strengthens the idea that this element becomes greatly preponderating farther down, in the mantle. The fact that the so-called plateau basalts, which have issued in vast thick floods from fissures in India, Oregon, Washington and elsewhere, are exceptionally high in iron oxides[133] indicates that they come up from the deeper portions of the crust. Their issuance from extensive fissures and in vast floods, covering hundreds of thousands of square miles and aggregating many thousands of feet in thickness, points the same way.

Metallogenic elements.—The principle of gravitative adjustment, the small amount of metallogenic elements in the rocks of the epiperisphere and some other considerations[4] have led to the idea that within the iron core there may be a central mass made up of metallogenic metals (which are mostly very heavy), such as copper, silver, gold, zinc, lead, the platinum metals and other rarer ones and their compounds with sulphur, selenium, arsenic, antimony.

The absence of the metallogenic elements, except traces of copper and the platinum metals, from meteorites, as shown by Merrill's researches, is opposed to this idea, unless the absence is explicable by the probable fact that meteorites are fragments of a cosmic body much smaller than the earth. The fact that the metallogenic metals can alloy or be miscible with nickel and iron is also against this view, but their sulphides, arsenides and other such compounds, being wholly or almost wholly immiscible with metals, might form such a core.

An alternative suggestion is that the metallogenic elements are abundant in the lower, lithosporic zone of the mantle; but here, again, the evidence of meteorites is adverse. The mineral relations of some of the metallogenic elements of lesser atomic weight and number, notably copper, zinc, silver and tin, to the transitional iron triad and to the igneous rocks[8] suggest that these metallogenic elements are most abundant in the upper part of the mantle, *i.e.*, in the shell immediately below the crust.

But, in dealing with the distribution of the metallogenic elements within the earth we are confronted with a paucity of data and with the inaccessibility of the regions to be considered, so that little more than speculative suggestions can be made.

Radioactivity.—A factor that has great influence on the observed thermal gradients, as also on speculations regarding the age of the earth, is the presence and the distribution of radioactive substances, especially uranium, thorium and radium.[134] With their great heat-giving power, it is clear that the presence of these elements in rocks will greatly modify the thermal gradients due to conduction and cooling alone, and they have even been invoked (as by Dutton) to explain volcanism. To what depth they occur is somewhat uncertain, but many researches, especially by Strutt, Joly and Holmes, are convergent and concordant in showing that the radioactive elements are more abundant in the rocks with much silica and alkalies (such as granite, syenite, trachyte and phonolite) than they are in rocks that are low in these constituents (such as gabbro, basalt and peridotite). The rocks that show the maximum amounts of radium (observed by Joly) are the leucitic lavas of Vesuvius, which are very high in potash, a correlation that it is tempting to think is connected with the relative richness of the highly potassic rocks in barium, an element that belongs to the same periodic group as radium. Strutt, Joly and Holmes conclude that by far the greater part of uranium, radium and thorium is confined to the uppermost portion of the crust. These radioactive elements, including uranium, are petrogenic[8] and consequently are most abundant in the perispheric igneous rocks and are probably very

rare or wanting in those of the interior, especially in the core. The facts that "the ultra-basic stony material of meteorites is poorer in radium than are terrestrial ultra-basic rocks" and that the nickel iron of meteorites and the terrestrial iron of iron-bearing basalts are free from radium[141,146] are certainly "significant," as Holmes says. Taken in connection with the modern view that the core is wholly or mostly of nickel iron, similar to that of meteorites, the distribution strengthens the view that the radioactive elements are confined almost wholly to the crust and that they are petrogenic. The influence of the presence of radioactive elements on the temperature of the earth at moderate depths will be discussed in Chap. VI. The bearing of their presence and distribution on the age of the earth has been dealt with by several writers;[142,143,145] it need not be discussed here. But the reader may be referred to Soddy's interesting speculations on the fate of the earth as determined by the distribution of radium,[144] the percentage of which in the crust has been estimated to be only 0.000,000,000,00x.[8]

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CHAPTER VI

OBSERVED TEMPERATURES IN THE EARTH'S CRUST*

C. E. Van Orstrand

(*Geophysicist, U. S. Geological Survey*)

Thermal constants for a few deep mines and about 200 deep wells are summarized in Tables 18 to 21. The records were selected from a limited number of mines and about 1,000 deep wells distributed throughout the globe.

MAJOR SOURCES OF HEAT

Several hypotheses have been proposed to explain the origin of the earth's internal heat, but at present only two fundamental heat sources are postulated—radioactivity and gravitational contraction (see Chap. VII). In addition to these hypotheses, a new hypothesis is now offered in which it is assumed that the earth's heat is derived in various proportions from both of the postulated sources. A vast amount of research has been devoted to this subject, but the fact remains that the origin and maintenance of the earth's internal heat continue to be one of the outstanding unsolved problems of science.

MINOR SOURCES OF HEAT

The development of minor quantities of heat in the earth's crust has been attributed to chemical reactions, friction on fault planes, porous plug expansion, heat developed by compression and shear and many other causes.

With regard to these minor heat sources, Nutting[43] has shown by the application of theoretical deduction to experimental data that shear and compression are not of much importance as heat sources; and some recent observations by the present writer in a deep well at Long Beach, Calif., suggest the conclusion that the temperatures in some faulted zones may decrease slightly rather than increase as the fault is approached. The increased conductivity of the rocks resulting from an increase in the moisture content of the fractured relative to the unfractured rocks lends support to this point of view.

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Chemical reactions undoubtedly cause marked departures from true rock temperatures in some mines; and a moderate generation of heat in deep-seated beds will ultimately produce an appreciable change in the temperature gradients.[49]

Important calculations by Dr. L. H. Adams[50] show that water rising along a crack in the rock from a depth of $3\frac{1}{2}$ km., corresponding to a release of pressure of about 1,000 megabars, will be subjected to an increase in temperature of more than 20°C. while in transit to the surface of the ground.

A further consideration of minor heat sources seems unnecessary for the purposes of this paper, but, because of its importance in terrestrial phenomena, one other minor heat source—solar heat—will be considered in a subsequent section.

THE EARTH'S ALBEDO AND EMISSIVITY

The albedo of the earth is 0.29. This means that approximately 29 per cent of the heat that reaches the earth from the sun is reflected back into interstellar space and that the remaining 71 per cent is absorbed by the earth. The albedo of the moon is only 0.07. Thus, the moon absorbs nearly all of the heat that it receives from the sun during the period of 14 days of sunshine, and subsequently, during equally long intervals of lunar nights, the heat thus absorbed is radiated back into interstellar space.

The earth's albedo and emissivity vary from place to place.[51] Forests and vegetation are good reflectors of heat, and, besides, the leaves of plants and trees absorb enormous quantities of thermal energy in the process of growth. As bare earth, on the other hand, is a good absorber and a poor radiator, it follows that the annual mean temperature of the ground in the forests is lower than it is in the adjacent area void of vegetation.

Callender[52] has shown that snow increases and cold rain lowers the soil temperatures with reference to the temperatures of the superposed atmosphere.

Additional evidence on the dependence of soil temperature on the moisture in the rocks is shown in Fig. 3 in which the upper curve represents the temperatures in a deep well located on land at a distance of about 300 ft. from the shore line of the Pacific Ocean. The lower curve represents the temperatures in a well located in the ocean at a distance of about 600 ft. from the shore line and on a line passing through the first well and at right angles to the shore line. The point m represents the annual mean temperature of the soil just beneath the surface of the ground, and l represents a similar point just beneath the

floor of the ocean. *km*, the excess of the annual mean soil temperature just beneath the surface of the ground over the annual mean air temperature just above the surface of the ground, has been designated by *e*, values of which are given in Tables 18 and 19.

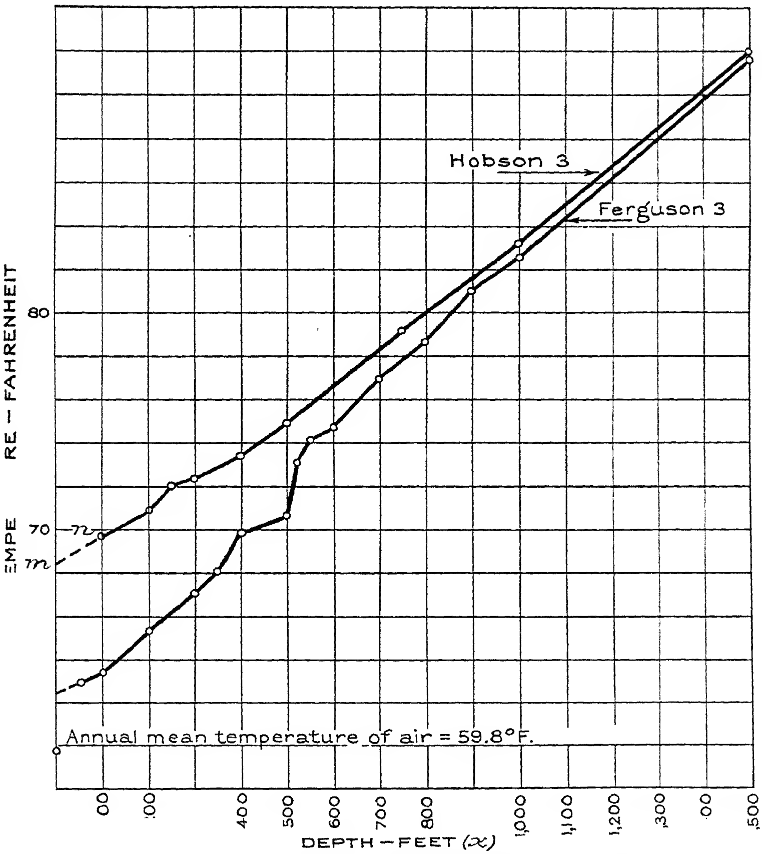


FIG. 3.—Rock temperatures beneath air (upper curve) and ocean floor (lower curve) at Ventura, Calif. The flat segment between 400 and 500 ft. probably represents convection in a water bed beneath the ocean.

The average value of *e* for 144 stations in the oil fields of southern California is 8.44°F. (4.69°C.). The same for 514 stations distributed chiefly in the oil fields of the United States, exclusive of California, is 1.54°F. (0.86°C.). Comparison of the values 4.69 and 0.86°C. suggests the possibility that *e* is dependent largely on rainfall and that the

absence of rainfall in southern California tends to increase the value of e. Furthermore, comparison of km with kl in Fig. 3 shows the same result, as does also the low value of 0.6°C. for e at Seal Beach (Table 18) in a well located near the shore line of the ocean and at an elevation of

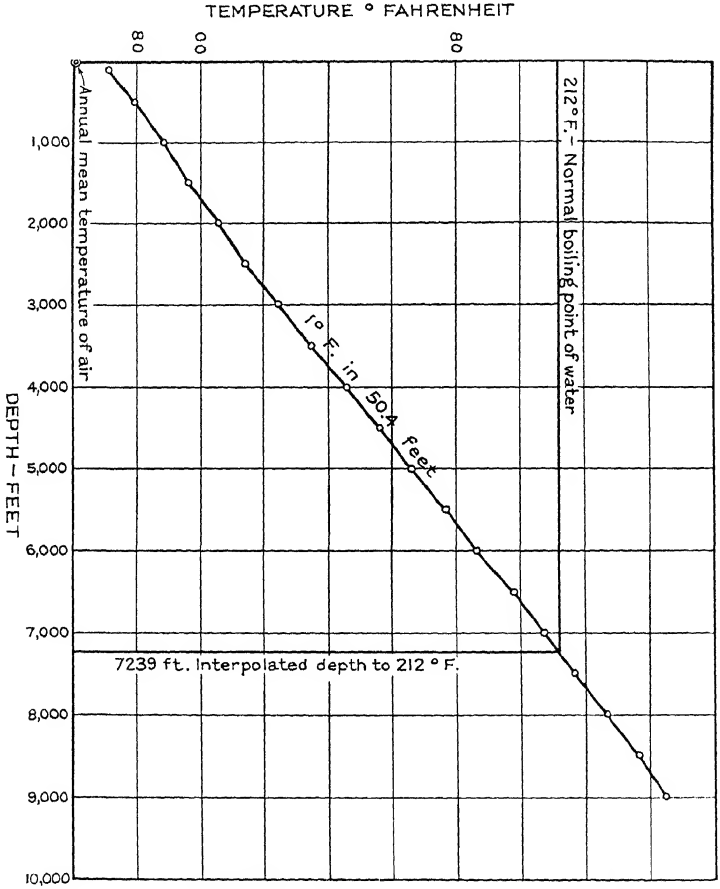


FIG. 4.—Depth-temperature curve of deep well, Nesa 11, Long Beach, Calif.

only 8 or 10 ft. above sea level so that the soil here is always saturated with water.

From the preceding discussion, it follows that the absorption of heat at the surface of the earth is dependent on its albedo; the loss, on its emissivity. The loss by radiation is equal to thermal conductivity (k) $\times$ thermal gradient (dv/dx) as determined from the depth tem-

perature curves, Figs. 3 and 4: and this is equal to the emissivity (E) $\times$ the temperature difference, which in this case is equal to e. The loss of heat by radiation is proportional therefore to e.

PERIODIC FLOW OF HEAT INTO AND OUT OF THE EARTH

Observations show that solar heat in temperate zones does not penetrate beneath level surfaces of the earth to depths exceeding 50 or 100 ft. The temperature at these shallow depths oscillates above and below the line mn, Fig. 3, which represents an extension of the observed depth temperature curve to the surface of the ground. The line mn having been determined, the temperature at any depth x and time t is found by adding to or subtracting from the particular ordinate of mn the corresponding value of v_p computed from the equation[53,54,55]

$$v_p = v_0 e^{-\frac{x}{\sqrt{\kappa}}\sqrt{\frac{\pi}{T}}} \sin\left(\frac{2\pi t}{T} - \frac{x}{\sqrt{\kappa}}\sqrt{\frac{\pi}{T}}\right) \tag{11}$$

where v_0 = half range of the soil temperature at zero depth.
κ = diffusivity of the soil.
T = period = 1 year = 31,556,927 sec.

From Eq. (11), the amplitude A at depth x is

$$A = v_0 e \qquad = v_0 \times \text{exponential factor.} \tag{12}$$

Table 15 contains values of the exponential factor in (12) for different values of x and κ. The coefficients of diffusivity have been taken from the volume by Ingersoll and Zobel.[53]

TABLE 15

Exponential Factors for Values of x and κ
(Metric units)

Depth (x)		κ			
Ft.	m.	0.0031 dry soil	0.0049 damp soil	0.0064 crustal rocks	0.0133 sandstone
1	0.30	0.8414	0.8716	0.8867	0.9200
5	1.52	0.4216	0.5031	0.5482	0.6591
10	3.05	0.1778	0.2531	0.3005	0.4343
15	4.57	0.0749	0.1273	0.1648	0.2863
20	6.10	0.0316	0.0641	0.0903	0.1887
25	7.62	0.0133	0.0322	0.0495	0.1243
30	9.14	0.0056	0.0162	0.0271	0.0819
35	10.67	0.0024	0.0082	0.0149	0.0540

Table 16 contains the results of a few observations in the United States from which values of the other factor (v_0) in Eq. (12) can be computed. The required values are obtained by dividing the tabular values in the third from the last column by 2.

TABLE 16

PRECIPITATION, ANNUAL MEAN TEMPERATURE OF AIR AND SOIL AND RANGE OF MEAN MONTHLY TEMPERATURES IN AIR AND SOIL

Location		Depth		Annual mean temperature,			Range of mean monthly tempera- tures, °C.		Precipita- tion	
Town	State	In.	Cm.	Air	Soil	e	Air	Soil	In.	Cm.
Auburn	Ala.	3.0	7.6	17.6	18.3	−0.7	19.1	21.1	51.86	131.7
Davis	Calif.	0.5	1.3	16.4			17.1	23.7	16.62	42.2
Lincoln	Neb.	1.0	2.5	12.7	14.6	−1.9	32.0	34.8	27.94	71.0
Pendleton	Ore.	4.0	10.2	11.2			.6	32.2	13.79	35.0
Temple	Tex.	1.0	2.5	20.7	21.9	-1.2	20.6	22.8	33.96	86.3
Urbana	Ill.	1.0	2.5	10.6	11.8	-1.2	27.8	27.2	35.37	89.8

The differences between the annual mean temperatures of the air and the soil at the shallow depths given in the table correspond in magnitude and sign to the values of e in Tables 18 and 19. At Davis and Pendleton, where the rainfall[56] is small, the ranges in the soil temperatures are much greater than in the air temperatures; and even

TABLE 17

TEMPERATURE LAG

(t days)

Depth		κ			
Ft.	m.	0.0031 dry soil	0.0049 damp soil	0.0064 crustal rocks	0.0133 sandstone
1	0.30	10.0	8.0	7.0	4.8
5	1.52	50.2	39.9	34.9	24.2
10	3.05	100.4	79.9	69.9	48.5
15	4.57	150.6	119.8	104.8	72.7
20	6.10	200.8	159.7	139.8	96.9
25	7.62	251.0	199.7	174.7	121.2
30	9.14	301.2	239.6	209.6	145.4
35	10.67	351.4	279.5	244.6	169.7

in areas of large rainfall, except at Urbana, the ranges in the soil temperatures are slightly greater than in the air.

Again referring to Eq. (11), we have for the lag in time at which maximum or minimum temperatures reach a given depth

$$t = \frac{x}{2\sqrt{\kappa}}\sqrt{\frac{T}{\pi}}. \tag{13}$$

Values of t computed from Eq. (13) are given in Table 17.

Depths at which the lag is 180 days experience a reversal of seasons. At these depths, the highest temperatures occur in the winter; the lowest in the summer.

OBSERVED TEMPERATURES BENEATH THE INFLUENCE OF SOLAR HEAT

In making temperature surveys of deep wells, the present writer has adopted the practice of beginning the survey at a depth of 100 ft. so as to avoid the effects of seasonal changes and at the same time to permit of making a fairly accurate estimate of the annual mean temperature of the soil at zero depth. As the complete survey includes readings at intervals of either 100, 250 or 500 ft., it is possible to make a least-square adjustment of the straight line

$$v = a + bx \tag{14}$$

over the first 1,000 ft. The value of a thus determined is assumed to be the annual mean temperature of the soil, and the difference between this value and the annual mean temperature of the air[56] at the location is the value of e tabulated in Tables 18 and 19. The values of $1/b$ are given for the same range of depth, 1,000 ft. The other values of a and b in the table are determined from all of the observations in the well. The significance of a in this case is merely that of an intercept on the temperature axis.

The values of $1/b$ in the last column are frequently smaller than the corresponding values in the first column. Hence, this type of curve is convex toward the depth axis, as illustrated by the depth-temperature curve of the Long Beach well, Fig. 4, and the corresponding sequence of reciprocal gradients 29.1, 27.7, 27.2 recorded in Tables 18 and 20.

In a summary[57] of the data for 679 wells, it was found that 59 per cent of the curves were convex, 36 per cent concave and only 5 per cent straight lines as required on the basis that the earth is a large homogeneous sphere which is radiating heat into interstellar space.

TABLE 18

Temperature Gradients and Related Constants in the United States
(Metric units)

Location					30.48–304.80 m.		30.48 m. to greatest depth				
Town or field[47]	County			Elevation, m.	e, °C.	$1/b$, m./°C.	Depth, m.	a, °C.	b, °C./m.	$1/b$, m./°C.	r, °C.
	Section	Town	Range								
Alabama											
Albany.....	6	6S	4W		−0.3	211.9	610	16.13	0.00725	137.8	0.21
Atwood.....	26	8S	14W	170	+1.6	52.0	762	15.24	0.01797	55.7	0.18
Birmingham	28	18S	2W		+1.7	60.0	610	15.49	0.01741	57.4	0.09
Fayette.....	Fayette				+0.9	70.4	381	15.72	0.01516	65.9	0.15
Vernon.....	22	15S	16W	99	+0.2	60.0	1,067	16.26	0.01735	57.6	0.20
Arkansas											
Eldorado...	4	19S	15W		+0.9	29.1	658	15.90	0.04212	23.8	0.57
Smackover.	8	16S	15W		+1.6	30.4	610	15.24	0.04105	24.4	0.62
California											
Alamitos Heights.	Los Angeles				−5.3	24.8	1,372	22.13	0.03848	26.0	0.19
Bakersfield.......	11	28S	27E	287	−6.9	45.8	1,295	21.52	0.02860	34.9	1.31
Bakersfield.......	30	28S	21E	213	−6.8	32.4	2,743	26.71	0.03439	29.1	1.80
Coalinga.........	12	20S	15E		−4.7	26.9	1,317	21.76	0.03580	27.9	0.01
Fullerton.........	Orange				−5.2	27.3	1,301	22.66	0.03498	28.6	0.41
Grass Valley[41]....	Nevada			671	+0.1	92.5	1,036	12.57	0 00961	104.1	0.14
Huntington Beach	Orange				−3.6	22.6	1,219	19.19	0 04930	20.3	0.50
Kettleman Hills..	3	22S	17E		−3.7	28.5	802	21.69	0 03283	30.4	0.34
Long Beach......	Los Angeles			103	−4.9	29.1	2,743	19.60	0 03618	27.7	0.53
Santa Fe Springs.	Los Angeles			47	−4.7	28.9	1,379	22.04	0.03507	28.5	0.26
Seal Beach.......	Los Angeles			3	−0.6	18.6	1,067	19.27	0.04112	24.3	0.81
Ventura..........	Ventura			17	−4.5	36.6	1,067	19.66	0.02980	33.6	0.16
Whittier.........	19	3S	10W	137		32.9	1,143	22.83	0.03416	29.3	0.35
Colorado											
Calhan..........	9	12S	62W		-3.3	34.8	1,219	10.06	0.03120	32.0	0.59
Florence.........	32	18S	69W		-1.4	28.0	1,067	12.38	0.03921	25.5	0.57
Fort Collins......	8	9	68W		-3.4	32.9	1,219	10.23	0.03565	28.0	0.54
Longmont........	34	2N	69W		-4.7	30.9	1,981	10.41	0.04373	22.9	1.21
Illinois											
Bridgeport..	27	3N	12W		-0.2	30.0	549	12.71	0.03144	31.8	0.29
Casey......	7	11N	14W		-2.1	53.0		13.85	0.02092	47.8	0.19
Robinson...	9	7N	12W		-0.1		305	12.42	0.03589	27.9	0.16

TABLE 18.—(*Continued*)

Location					30.48–304.80 m.		30.48 m. to greatest depth				
Town or field[47]	County			Elevation, m.	e, °C.	1/b, m./°C.	Depth, m.	a, °C.	b, °C./m.	1/b, m./°C.	r. °C.
	Section	Town	Range								
Iowa											
Ames[42]	Story			.....	−0.9	78.9	640	9.83	0.01294	77.3	0.18
Kansas											
Augusta	20	28S	4E		−0.8	27.0	686	14.62	0.03478	28.7	0.27
Eldorado	28	25S	5E	417	+0.6	28.1	686	12.93	0.03543	28.2	0.23
Florence	28	21S	5E		−0.4	31.7	655	13.82	0.03002	33.3	0.22
Haverhill	22	27S	5E		−0.5	27.9	686	14.03	0.03529	28.3	0.24
Syracuse[43]	5	26S	41W	1,055	−1.7	45.0	1,402	12.78	0.02603	38.4	0.72
Kentucky											
Prater Creek	Floyd				−0.4	66.9	457	14.23	0.01161	86.1	0.39
Rockcastle Creek	Martin			252	+0.8	65.0	457	12.77	0.01393	71.8	0.33
Louisiana											
Blue Lake	14	7N	14W		−1.7	30.4	610	18.77	0.04101	24.4	0.69
Caddo Lake*	8	20N	16W		+0.6	20.7	610	18.79	0.04416	22.7	0.38
Haynesville	18	23N	8W		+0.1	26.4	838	18.01	0.04092	24.4	0.44
Homer	29	21N	7W		+0.2	20.3	610	17.70	0.05426	18.4	0.49
Many	13	7N	11W		+0.1	28.5	762	16.94	0.04145	24.1	0.80
Monroe	21	16N	6E		−0.3	26.8	686	18.66	0.0401[illegible]	24.9	0.29
Pine Island	13	21N	15W		+0.2	19.3	1,067	20.63	0.04043	24.[illegible]	0.80
Sligo	14	17N	12W		+0.4	21.2	503	19.02	0.04225	23.6	0.40
Zwolle	10	8N	13W		−2.7	27.7	1,030	20.57	0.03994	25.1	0.42
Michigan											
Houghton	1	53N	35W	.....	−0.7	71.2	504	5.22	0.01429	70.0	0.05
Keweenaw Point†	Keweenaw			.....	−1.7	54.4	1,636	6.36	0.01553	64.4	0.73
Lake Linden	6	55N	32W	.....	−2.4	87.9	437	6.19	0.01380	72.5	0.19
Surrey Tp	Clare			.....	0.0	46.3	1,128	4.52	0.03439	29.1	1.22
Mississippi											
Amory	4	13S	18W	126	−0.1	89.1	762	15.97	0.01595	62.7	
Monroe Co.	22	15S	17W	71	−0.4	101.2	533	17.00	0.01380	72.5	
Jackson‡	13	5N	1E	.....	−0.5	19.1	686	19.25	0.05226	19.1	0.41

TABLE 18.—(*Continued*)

Location				Elevation, m.	30.48–304.80 m.		30.48 m. to greatest depth				
Town or field[47]	County				e, °C.	$1/b$, m./°C.	Depth, m.	a, °C.	b, °C./m.	$1/b$, m./°C.	r, °C.
	Section	Town	Range								
Montana											
Anaconda[44]	Deerlodge				+1.2	20.8	488	5.89	0.03848	26.0	2.19
Conrad	32	30N	2W	1,097	−1.8	44.3	610	7.68	0.02227	44.9	0.06
Kevin-Sunburst ..	21	34N	1W	1,081	−2.4	60.4	1,247	8.42	0.01759	56.9	0.53
Nevada											
Virginia City[45]	Storey			.....	+0.3	16.1	701	10.48	0.05659	17.7	0.87
New Jersey											
Franklin Furnace	Sussex			.....	−0.2	151.5	762	10.44	0.01066	93.7	0.14
New Mexico											
Artesia	13	22S	29E	933	−2.8	98.0	881	20.48	0.00800	124.9	0.12
Carlsbad ...	24	20S	29E	1,009	−2.2	100.7	,829	17.43	0.01383	72.3	0.62
Lovington .	11	16S	32E	1,341	−0.9	43.8	,372	20.01	0.01010	99.0	1.03
Roswell	25	11S	27E	1,134	+3.3	88.1	914	18.61	0.01063	94.1	0.13
New York											
Greenwood	Steuben			546	−1.0	35.2	1,295	7.44	0.02931	34.1	0.57
Tyrone	Schuyler			376	−0.7	30.3	533	8.55	0.03476	28.7	0.13
Willing Tp.	Allegany			638	−1.2	44.1	1,433	5.18	0.03093	32.3	1.09
North Dakota											
Lonetree ...	9	155N	85W	600	−3.4	34.8	1,143	4.44	0.03642	27.4	1.15
Oklahoma											
Ardmore		4S	2W		−0.5	66.0	914	16.96	0.01783	56.1	0.56
Ard-Healdton		3S	2W		0.0	59.0	838	17.17	0.01666	60.0	0.13
Bessie		10N	18W		−1.	77.6	457	17.16	0.01509	66.2	0.16
Billings	20	24N	1W		−0.8	29.0	1,067	15.64	0.03483	28.7	0.26
Blackwell	9	27N	1W		+0.7	26.9	610	14.48	0.03695	27.0	0.19
Braman	8	28N	1W		−1.3	28.3	719	16.78	0.03312	30.2	0.19
Burbank	22	27N	5E		0.0	29.0	869	14.08	0.03988	25.1	0.54
Cromwell	10	10N	8E		+0.7	28.0	914	14.62	0.04015	24.9	0.36
Davenport ...	2	14N	5E		+0.2	62.2	1,029	14.82	0.02118	47.2	0.56
Dilworth	8	28N	1E	340	−0.2	29.9	1,029	14.71	0.03623	27.6	0.17

TABLE 18.—*(Continued)*

Location					30.48–304.80 m.	30.48 m. to greatest depth				
Town or field[47]		County	Elevation, m.		1/b, m./	Depth, m.			1/b, m./°C.	
Oklahoma—(*Continued*)										
Drumright.....		18N 7E	262	−2.0	40.1	701	15.94	0.03279	30.5	0.67
Glenn Pool.....		18N 11E	...	+0.7	20.4	509	15.04	0.05000	20.0	0.21
Holdenville[46]...		7N 10E	...	−1.0	29.0	1,585	16.32	0.03649	27.4	0.69
Hubbard.......		26N 2W	...	−0.7	26.6	875	15.72	0.03833	26.1	0.27
Morris.........		13N 14E	209	0.0	21.5	491	16.24	0.04473	22.3	0.28
Newkirk.......	15	27N 3E	...	−0.3	24.2	914	15.29	0.04054	24.7	0.50
Okemah.......	26	12N 10E	...	+0.4	23.2	838	15.88	0.04216	23.7	0.28
Oklahoma City.	12	11N 3W	...	−0.3	76.3	1,829	13.93	0.02058	48.6	0.68
Oklahoma City.	13	11N 3W	...	+1.3	.1	1,829	12.59	0.02116	47.3	0.63
Papoose.......	2	9N 9E	...	+0.2	27.9	914	15.09	0.03977	25.1	0.46
Perry..........	9	20N 2W	...	+0.7	30.2	914	14.98	0.03130	31.9	0.23
Sasakwa.......	7	6N 8E	...	−0.6	33.4	1,067	15.60	0.03423	29.2	0.35
Seminole.......	18	9N 6E	...	−0.6	49.9	1,219	14.06	0.03060	32.6	1.20
Tonkawa......	2	24N 1W	...	−0.7	29.2	1,219	15.17	0.03676	27.2	0.55
Walters........	4	2S 10W	...	−0.8	73.2	655	17.16	0.01575	63.5	0.21
Wewoka.......	28	8N 8E	252	+0.4	27.3	914	15.48	0.03828	26.1	0.21
Oregon										
Astoria........	25	8N 10W	.....	−0.5	36.5	1,152	10.59	0.03044	32.9	0.20
Bonanza.......	19	39S 12E	.....	−0.2	9.6	1,180	−44.11	0.14826		1.08
Lakeview......		Lake	.....	−1.9	16.1	533	3.74	0.10369		3.37
Medford.......		Jackson	.....	−2.9	.....	274	14.80	0.02406	41.5	0.16
Vale...........	19	19S 44E	.....	−4.8	10.8	395	15.47	0.08339	12.0	1.18
Pennsylvania										
Gaines Junction.......		Tioga	.....	−1.0	51.7	1,676	5.02	0.02951	33.9	1.22
Genesee Tp...........		Potter	642	−0.5	40.6	1,524	5.09	0.02986	33.5	1.03
Hebron Tp...........		Potter	657	+1.3	35.2	1,600	3.84	0.03099	32.3	0.74
Johnstown...........		Westmoreland	829	−0.4	59.7	1,981	5.94	0.02393	41.8	1.23
Long Bridge..........		Westmoreland	344	+1.1	39.4	2,077	8.27	0.02853	35.1	0.79
New Castle...........		Butler	.....	+0.6	47.0	1,372	7.05	0.02955	33.9	1.08
Texas										
Big Lake....		Upton	879	−0.1	54.9	968	19.09	0.01274	78.5	0.54
Blue Ridge..		Fort Bend		−0.9	32.2	884	21.09	0.03281	30.5	0.30
Boggy Creek		Anderson		−0.9	21.9	1,129	18.25	0.05226	19.1	1.26
Bonham....		Fannin		−1.1	24.1	347	19.03	0.03984	25.1	0.26
Brownwood.		Brown	483	−2.3	33.5	381	20.43	0.02960	33.8	0.22

TABLE 18.—(*Continued*)

Location				Elevation, m.	30.48–304.80 m.		30.48 m. to greatest depth				
Town or field[47]	County: Section	Town	Range		e, °C.	1/b, m./°C.	Depth, m.	a, °C.	b, °C./m.	1/b, m./°C.	r, °C.
Texas—(*Continued*)											
Callisburg.	Cooke				−0.3	56.5	1,122	17.86	0.01790	55.9	0.23
Canadian..	Hemphill				−1.3	69.1	1,450	15.88	0.01825	54.8	0.65
Dale.......	Caldwell				−0.6	25.6	686	20.20	0.04190	23.9	0.20
Damon....	Brazoria				−2.7	38.2	1,219	22.02	0.02951	33.9	0.57
Del Rio....	Val Verde				−2.4	43.5	1,123	23.39	0.02390	41.9	0.42
Dickens..	Dickens				−1.9	53.6	732	18.57	0.01598	62.5	0.19
Eastland.	Eastland				−2.1	31.2	914	18.85	0.03436	29.1	0.26
Edwards.	Val Verde			539	−1.1	43.8	2,012	18.34	0.03037	32.9	1.74
Electra...	Wichita				−0.4	47.3	558	18.66	0.02231	44.8	0.11
Ennis.....	Ellis				−3.0	32.0	404	21.16	0.03383	29.6	0.32
Ferris.....	Ellis				−0.6	27.0	381	19.19	0.03689	27.1	0.36
Graford...	Palo Pinto			330	−1.8	31.6	914	19.62	0.03383	29.6	0.42
Graham...	Young				−1.2	30.4	1,321	19.09	0.03157	31.7	0.50
Greenville.	Hunt				−0.9	28.1	533	19.52	0.03381	29.6	0.22
Guthrie....	King				−1.7	51.4	899	19.11	0.01637	61.1	0.59
Humble........	Harris				−0.8	23.5	743	20.64	0.04633	21.6	0.43
Leonard.......	Fannin				−1.3	30.5	762	19.92	0.02710	36.9	0.54
Long Point.....	Fort Bend				−1.2	30.7	1,006	22.04	0.03217	31.1	0.26
Luling.........	Caldwell			136	−0.3	23.8	686	20.96	0.04187	23.9	0.28
Lytton Springs.	Caldwell				−0.9	21.8	507	21.86	0.04278	23.4	0.21
Mexia.....	Limestone				−0.4	25.1	914	18.98	0.04185	23.9	0.20
Odessa.....	Midland				−2.8	79.1	605	19.96	0.01307	76.5	0.12
Ozona.....	Crockett				−0.5	53.3	1,676	15.66	0.02960	33.8	1.13
Pampa....	Roberts			824	−1.1	50.8	1,524	16.44	0.01544	64.7	0.41
Panhandle.	Carson			995	−3.3	51.6	602	16.90	0.01950	51.2	0.61
Panhandle.....	Gray			988	−1.7	65.3	1,044	16.14	0.01228	81.4	0.18
Panhandle.....	Lipscomb			....	−1.1	67.9	1,277	16.29	0.01518	65.9	0.37
Pierce Junction.	Harris			18	−1.9	32.4	1,006	22.50	0.03195	31.3	0.20
Ranger........	Eastland				−1.7	26.9	914	18.58	0.03851	26.0	0.33
Sherman.......	Grayson				−0.6	37.6	442	18.54	0.02530	39.[illegible]	0.29
Smith's Ranch.	De Witt				−0.9	31.3	600	21.60	0.03467	28.9	0.30
St. Jo........	Montague			311	+1.0	47.3	527	16.18	0.02169	46.1	0.09
Waco.........	McLennan				−3.2	27.9	607	23.63	0.03161	31.7	0.31
Wolfe City....	Hunt				+0.2	28.7	610	17.32	0.03682	27.2	0.23
Utah											
Grand County.	33	22S	22E	...	−2.8	73.7	1,463	14.05	0.01839	54.4	0.57

TABLE 18.—(*Continued*)

Location					30.48–304.80 m.		30.48 m. to greatest depth				
Town or field[47]	County: Section	Town	Range	Elevation, m.	e, °C.	1/b, m./°C.	Depth, m.	a, °C.	b, °C./m.	1/b, m./°C.	r, °C.
Washington											
Benton City	27	11N	26W	.....	−3.1	24.2	671	15.24	0.03762	26.6	0.23
Moclips	8	20N	12W	50	+1.7	55.7	1,067	6.87	0.02512	39.8	0.44
Seattle	NW cor. 54th 32sd			15	+1.0	59.1	762	9.42	0.01730	57.8	0.13
West Virginia											
Bridgeport	Harrison			354	−1.3	47.0	2,228	9.86	0.02554	39.2	1.65
Fairmont	Marion			366	−0.8	44.1	2,286	9.58	0.02772	36.1	1.35
Grantsville	Calhoun			.....	0.0	47.6	1,372	11.51	0.02195	45.6	0.42
Spencer	Gilmer			267	+0.4	44.3	991	11.73	0.02149	46.5	0.12
Volcano	Wood			371	−0.4	63.8	1,341	9.46	0.02690	37.2	1.10
West Union	Doddridge			347	+0.3	51.6	1,981	7.63	0.02524	39.6	1.52
Wyatt	Harrison			.....	−0.8	50.5	610	12.01	0.01855	53.9	0.09
Wyoming											
Big Muddy	Converse			.....	−1.9	29.1	991	9.93	0.03434	29.1	0.29
Cody	24	51N	101W	1,649	−4.7	34.1	1,295	10.84	0.03095	32.3	0.41
Grass Creek	Hot Springs			.....	−1.2	27.4	332	8.47	0.03560	28.1	0.16
Lance Creek	2	35N	65W	.....	−3.6	22.2	1,067	9.68	0.04958	20.2	0.29
Lost Soldier	10	26N	90W	.....	−1.3	17.3	1,036	9.37	0.04682	21.3	1.14
Pine Mountain	Natrona			.....	+0.3	21.5	683	9.44	0.03749	26.7	0.95
Rawlins	5	25N	86W	2,118	−1.1	38.0	914		0.03192	31.3	0.64
Rock River	2	19N	78W	.....	−2.1	40.8	914	7.94	0.02608	38.4	0.23
Salt Creek	15	40N	79W	1,510	−2.1	34.2	808	8.79	0.03789	26.4	0.57
Teapot	28	39N	78W	.....	−1.6	24.6	834	8.03	0.05001	20.0	0.64
Thermopolis	35	43N	94W	.....	−3.6	.....	274	10.53	0.07666	13.1	0.32
Hot Springs Co.	23	44N	95W	.....	−4.1	30.9	429	10. [illegible]0	0.03545	28.2	0.38
Little Sand Draw	2	44N	96W	1,461	−2.7	37.6	914	8.48	0 02938	34.0	0.32

* Well in lake.
† 70 observations in mines.
‡ Buried volcano.

The constants *a* and *b* in Tables 18 and 19 are to be used in interpolating temperatures from 100 ft. to the greatest depth in the well (column 6). For extension of the curves beyond the greatest depths, the constants in Tables 20 and 21 are probably to be preferred in most cases, for these constants are based on observations from 2,000 ft.

TABLE 19

TEMPERATURE GRADIENTS AND RELATED CONSTANTS IN FOREIGN COUNTRIES
(Metric units)*

Town or section of country	Range of depth, m.			1/b, m./	Depth, m.		b, °C./m.	1/b, m./	
Cape of Good Hope									
Carnarvon[1].	30	306	−5.4	18.8	1,525	26.49	0.03088	32.4	1.78
Transvaal									
Witwatersrand[2]..	151	305	+0.7	66.4	1,194	16.43	0.00942	106.2	1.37
Witwatersrand..	305	610	−0.9		892	18.32	0.00817	122.6	0.31
Witwatersrand[3].	1,804	2,143	−2.3	91.5	2,143	18.48	0.00824	121.5	0.07
Borneo									
Samarinda[4].	233	387	+3.3		387	22.46	0.06363	15.7	0.30
Samarinda..	290	471	−7.7		471	33.39	0.03901	25.6	1.66
Samarinda..	420	574	+2.8		574	22.91	0.05330	18.8	2.14
Japan									
Echigo[5].	43	304	+1.3	24.2	607	11.26	0.04210	23.8	0.26
Echigo..	113	343	−2.8	34.7	726	11.13	0.04364	22.9	2.16
Echigo..	32	328	−0.4	23.6	611	11.98	0.05014	20.0	0.41
Australia									
Great Basin[6]......	219	544	−0.9	26.8	914	24.52	0.02827	35.4	2.44
New South Wales[7].	185	320	−1.5	84.9			0.02344	42.7	0.39
Northwestern[8].....	155	277	+4.1	10.6	770	31.14	0.03496	28.6	.72
South[9]............	415	649	+4.9	13.1	1,433	30.23	0.05012	20.0	.01
South[10]...........	94	344	+3.0	9.2	479	24.17	0.05610	17.8	4.40
Southwest[11].	93	450	−1.1	22.8	563	22.83	0.02851	35.1	2.69
Austria-Hungary									
Příbram[12]..		360	−1.6	60.1		8.98	0.01489	67.2	0.27
Příbram[13]..		369	−0.9	53.2	579	8.47	0.01520	65.8	0.59
France									
Martincourt[14]......................	600	1,200	−7.6	.....	1,200	16.61	0.02599	38.5	0.44
Pas-de-Calais[15].....................	130	1,364	−1.0	.....	1,364	11.01	0.01795	55.7	0.25
Rouchamp[16].........................	300	600	−1.1	29.7	1,009	8.86	0.03784	26.4	0.37
Germany									
Paruschowitz[17].	37	316	−4.6	43.2	1,959	10.49	0.02916	34.3	0.63
Pechelbronn[18]..	36	305	−5.3	20.8	516	13.04	0.06381	15.7	1.04
Schladebach[19]..	36	306	−1.6	36.7	1,716	9.53	0.02801	35.7	0.41
Sennewitz[20]....	451	634	+0.7	42.4	1,084	9.42	0.02129	47.0	0.43
Sperenberg[21]...	220	596	−6.3	33.8	1,269	16.74	0.02641	37.9	0.84

TABLE 19.—(*Continued*)

Town or section of country	Range of depth, m.		e, °C.	1/b, m./°C.	Depth, m.	a, °C.	b, °C./m.	1/b, m./°C.	
				England					
Bristol[22]	123	539	−0.1	.....	539	9.32	0.02654	37.7	0.07
Ince[23]	147	614	−4.9	41.6	745	13.23	0.02741	36.5	0.56
Kentish Town[24]	30	305	−0.2	31.5	335	10.23	0.03212	31.1	0.72
				Wales					
Flint County[25]	142	317	−0.1	.....	317	9.02	0.02382	42.0	1.30
Glamorganshire[26]	166	388	−0.1		388	8.98	0.02637	37.9	0.06
				Netherlands					
	198	643	−2.7	33.9	1,000	11.77	0.03004	33.3	0.29
Sevenum[28]	100	400	−3.3	30.4	1,000	13.06	0.02949	33.9	0.84
Witwaterings[29]	300	700	+2.7	29.2	1,245	6.42	0.03434	29.1	0.54
Woensdrecht[30]	230	670	−3.4	36.4	1,200	9.06	0.03806	26.3	1.13
Utrecht[31]	35	365	−1.2	.....	365	10.52	0.01923	52.0	0.11
				Poland					
Bitkow[32]	50	300	+0.1	60.9	950	6.74	0.02052	48.7	0.32
Boryslaw	52	300	−1.2	53.7	1,500	7.41	0.02667	37.5	0.41
Boryslaw	50	300	−1.1	52.8	1,275	7.12	0.02687	37.2	0.47
Boryslaw	100	300	−1.5	53.7	1,350	8.04	0.02147	46.6	0.48
Krosno	50	300	−1.0	44.2	600	8.49	0.02357	42.5	0.16
Tustanowice	50	300	−2.7	59.9	1,200	7.73	0.02574	38.8	0.74
				Russia					
Bibi-Eibat[33]†	267	617	−5.2	32.2	986	19.41	0.03170	31.5	1.93
Surakhan	181	459	+2.1	21.3	600	15.19	0.03880	25.8	0.78
Donetz Basin[34]	95	295	−3.1	41.9	1,200	9.94	0.02891	34.6	0.61
Donetz Basin	85	306	−1.2	37.0	540	9.72	0.02732	36.6	0.17
Kharkov[35]	50	350	−1.7	37.8	565	12.13	0.01706	58.6	1.09
Kharkov	50	350	−3.4	68.1	450	12.44	0.01478	67.6	0.03
Moscow	100	300	−0.9	59.0	700	6.68	0.02009	49.8	0.3
				Canada					
Ontario[36]	76	345	−0.3	53.4	934	2.68	0.01112	90.0	0.58
Ontario	72	411	−4.2	123.0	1,186	4.95	0.00817	122.6	0.11
Ontario	55	374	−4.2	108.9	1,178	5.28	0.00782	127.9	0.18
Ontario	335	792	−2.1	73.4	1,288	3.43	0.01321	75.7	0.09
Ontario	119	607	−4.3	95.7	1,090	5.28	0.01092	91.7	0.10
Ontario	179	611	−2.2	71.5	915	37	0.01540	65.0	0.50
Ontario	151	604	−2.4	79.0	912	50	0.01281	78.0	0.13
Ontario	122	305	−3.7	65.8	945	.69	0.01174	85.1	0.38
Alberta[37]	76	305	−1.1	39.0	457	.39	0.02563	39.0	0.08
Alberta	76	305	−0.5	22.9	386	.22	0.04065	24.6	0.48
				Mexico					
Túxpam[38]	30	818	+0.8	.....	818	23.39	0.05570	17.9	2.28
Túxpam	23	297	−3.1	25.5	1,244	24.43	0.04522	22.1	1.92
				Brazil					
Minas Geraes[39]	1,228	1,780	+0.6	67.0	1,871	17.13	0.01551	64.5	0.31
				Colombia					
Puerto Wilches[40]	30	305	+1.2	66.1	945	25.96	0.01531	65.3	0.44

* For conversion factors, see Units in Appendix.

† One observation in each of 139 wells.

TABLE 20

Temperature Gradients in the United States Based on Observations from 2,000-Ft. (610 M.) to Greatest Depth in the Well
(Metric units)

Town or field		b, C./	1/b, m./°C	
Alabama				
Vernon............	15.91	0.01783	56.1	0.09
California				
Alamitos Heights..	22.56	0.03804	26.3	0.15
Bakersfield........	14.69	0.03518	28.4	0.99
Coalinga..........	21.41	0.03613	27.7	0.11
Fullerton..........	22.97	0.03474	28.8	0.42
Grass Valley......	13.39	0.00860	116.3	0.06
Huntington Beach...	16.81	0.05187	19.3	0.27
Long Beach.........	18.31	0.03684	27.2	.35
Santa Fe Springs....	23.47	0.03370	29.7	0.09
Seal Beach..........	19.77	0.04017	24.9	.07
Ventura............	19.78	0.02971	33.6	.04
Whittier.	23.79	0.03326	30.1	0.21
Colorado				
Calhan.......	6.33	0.03507	5	0.37
Florence......	8.46	0.04371	9	0.50
Fort Collins...	6.90	0.03911	6	0.30
Longmont....	6.76	0.04641	6	0.47
Kansas				
Syracuse...........	8.18	0.03017	33.1	0.32
Louisiana				
Haynesville.	21.31	0.03665	3	0.00
Pine Island..	24.59	0.03565	0	0.26
Zwolle......	22.52	0.03760	6	0.43
Michigan				
Houghton Baltic 16*.	2.38	0.01681	5	0.06
Keweenaw Point....	5.46	0.01609	2	0.71
Surrey Tp..........	−2.59	0.04261	5	0.56
Montana				
Kevin-Sunburst.....	7.28	0.01863	53.7	0.74
New Mexico				
Artesia.............	20.99	0 00727	137 7	0.06
Carlsbad...........	15.29	0.01538	65 0	0.43
Lovington..........	24.24	0.00603	165 5	0.18
Roswell............	17.97	0.01148	87 1	0.14
New York				
Greenwood Tp......	4.29	0.03241	30.9	0.16
Willing Tp..........	−0.09	0.03558	28.1	0.68
North Dakota				
Lonetree............	−2.30	0.04373	22.9	0.63
Oklahoma				
Ardmore...........	12.90	0.02315	2	0.83
Ardmore-Healdton..	17.01	0.01691	1	0.07
Billings............	15.59	0.03492	6	0.
Burbank...........	11.01	0.04422	6	0.43
Cromwell..........	11.41	0.04440	5	0.18
Davenport.....	10.81	0.02597	38.5	0.23
Dilworth......	14.33	0.03669	27.3	0.04
Holdenville....	14.01	0.03829	26.1	0.34
Hubbard......	12.79	0.04218	23.7	0.23
Newkirk......	11.87	0.04484	22.3	0.86
Oklahoma—(*Continued*)				
Okemah...........	14.70	0.04353	23.0	0.36
Oklahoma City......	11.64	0.02231	44.8	0.62
Oklahoma City......	10.77	0.02251	44.4	0.54
Papoose...........	11.54	0.04447	22.5	0.43
rry............	13.38	0.03336	30.0	0.00
Sasakwa...........	13.69	0 03653	27.4	0.11
Seminole...........	7.63	0 03744	26.7	0.52
Tonkawa...........	11.86	0.04017	24.9	0.51
Wewoka...........	13.73	0 04061	24.6	0.14
Oregon				
Astoria............	10.23	0.03089	32.4	0.07
Pennsylvania				
Gaines Junction.	−0.28	0.03377	29.6	0.36
Genessee Tp....	0.09	0.03412	29.3	0.61
Hebron Tp.........	0.74	0.03346	29.9	0.56
Johnstown.........	3.08	0.02590	38.6	1.22
Long Bridge.....	6.29	0.02975	33.6	0.73
New Castle.....	1.25	0.03507	28.5	0.25
Texas				
Big Lake.....	−3.23	0.03122	32.0	0.96
Blue Ridge....	18.68	0.03614	27.7	0.04
Boggy Creek..	13.89	0.05683	17.6	1.19
Callisburg....	16.55	0.01934	51.7	0.09
Canadian.....	13.09	0.02082	48.1	0.56
Damon.......	18.88	0.03281	30.5	0.25
Del Rio......	26.66	0.02040	49.0	0.11
Eastland......	18.93	0.03445	29.0	0.18
Edwards......	7.44	0.03713	26.9	0.48
Graford......	17.04	0.03737	26.8	0.18
Graham........	16.37	0 03407	29.4	0.53
uthrie.........	13.36	0 02353	42.5	0.85
Long Point......	23.21	0.03066	32.6	0.09
Mexia..........	19.82	0 04083	24.5	0.04
Ozona..........	12.02	0 03255	30.7	0.63
ampa.........	16.26	0.01557	64.2	0.52
Panhandle......	15.02	0.01356	73.7	0.17
Panhandle......	14.82	0.01664	60.1	0.36
Pierce Junction.	23.73	0.03040	32.9	0.06
Ranger.........	15.	.04283	23.4	0.28
Utah				
Grand County.......	12.15	0.02005	49.9	0.35
Washington				
Moclips............	4.82	0.02778	36.0	0.06
West Virginia				
Bridgeport....	4.88	0.02865	34.9	1.17
Fairmont.....	4.48	0.03095	32.3	1.24
Grantsville....	8.35	0.02535	39.4	0.26
Spencer.......	10.86	0.02260	44.2	0.08
Volcano......	3.94	0.03223	31.1	0.33
West Union...	2.47	0.02878	34.7	0.82
Wyoming				
Big Muddy......	8.22	0.03644	27.4	0.21
Cody...........	12.99	0.02893	34.6	0.29
Lance Creek.....	10.19	0.04903	20.4	0.08
Lost Soldier......	13.37	0.04159	24.0	0.26
Rawlins.........	0.65	0.03901	25.6	0.24
Rock River......	4.44	0.03080	32.5	0.24
Salt Creek.......	5.07	0.04371	22.9	0.39
Teapot.........	6.07	0.05322	18.8	0.32
Little Sand Draw.	5.82	0.03303	30.3	0.19

* From 4,815 to 6,250 ft.

TABLE 21

TEMPERATURE GRADIENTS IN FOREIGN COUNTRIES BASED ON OBSERVATIONS FROM ABOUT 610 M. TO GREATEST DEPTH IN THE WELL

(Metric units)

Town or section of country			1/b, m./	
Cape of Good Hope				
Carnarvon.	34.94	0.02320	43.1	0.27
Transvaal				
Witwatersrand..	12.69	0.01341	74.5	0.59
Witwatersrand..	20.17	0.00580	172.7	0.32
Australia				
Great Basin.	16.76	0.03866	25.8	1.49
South.......	42.63	0.03944	25.3	2.41
Austria-Hungary				
Přibram............	10.74	0.01258	79.5	0.21
France				
Rouchamp.	5.82	0.04123	24.2	0.24
Germany				
Paruschowitz........	9.11	0.03015	33.1	0.60
Schladebach.........	8.29	0.02900	34.5	0.40
Sennewitz...........	9.23	0.02145	46.6	0.40
Sperenberg..........	20.32	0.02293	43.[illegible]	
Netherlands				
Oploo........	9.72	0.03235	30.9	0.14
Sevenum.....	6.81	0.03691	27.1	0.98
Witwaterings.	5.53	0.03523	28.4	0.67
Woensdrecht..	5.07	0.04236	23.6	

Town or section of country				
Poland				
Bitkow.	4.49	0.02344	42.6	0.35
Boryslaw....	7.56	0.02659	37.6	0.35
Boryslaw....	6.09	0.02798	35.7	0.38
Boryslaw....	5.29	0.02402	41.6	0.39
Tustanowice.	3.16	0.03068	32.6	0.17
Russia				
Bibi-Eibat....	22.39	0.02752	36.3	1.49
Donetz Basin.	8.56	0.03068	32.6	0.49
Canada				
Ontario.............	4.09	0.00900	111.1	0.14
Ontario.	5.05	0.00804	124.3	0.12
Ontario.	4.92	0.00817	122.4	0.07
Ontario.	3.62	0.01303	76.8	0.10
Ontario.	5.15	0.01108	90.2	0.11
Ontario.............	5.24	0.01075	92.9	0.16
Ontario.............	3.78	0.01247	80.3	0.24
Ontario.............	9.30	0.00988	101.3	0.49
Mexico				
Tuxpam............	18.73	0.05122	19.	1.69
Colombia				
Puerto Wilches.	[illegible]3	0.01598	62.5	0.2

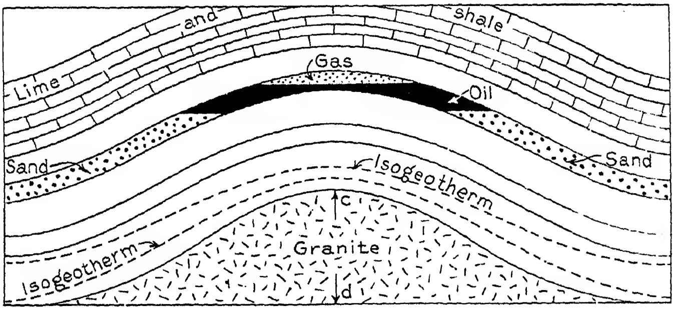
to the greatest depth in the well and are therefore practically free from the effects of surface topography and other surface irregularities. Nearly all of the constants are based on data from individual wells, and in the selection of the wells preference has been given to depth of well and reliability of data.

The quantity r is the probable error of an observation of weight unity. It must not be interpreted in the strict sense usually attributed

to it in geodesy and astronomy. On account of the curvature of the depth-temperature curves (Figs. 3 and 4), the values of r may represent chiefly deviation from linearity, and, in other cases, particularly in mines and in those oil fields in which single observations have been made in each well of a group, a large value of r may imply that the observed values are widely and irregularly scattered.

TEMPERATURE GRADIENTS IN SEDIMENTARY STRATA

As explained in the preceding section, temperature gradients vary in the vertical, and it is sometimes also possible to detect a variation in the horizontal over both local and regional areas.



RECIPROCAL GRADIENTS, FEET PER °F.				
	100-1000 FEET	100-2000 FEET	100-3000 FEET	100-4000 FEET
CREST	49.8	50.6	50.0	47.4
FLANK	62.1	57.3	56.0	51.3

FIG. 5.—Variation of reciprocal gradients in the horizontal.

Figure 5 is a cross section of a typical oil field showing the sedimentary strata superimposed over a granite ridge or dome. The reciprocals of the averages of the gradients in 57 oil fields are given as accurately as possible for the crests and the flanks of such structures. The tabulation shows that, on the crests of the structures, the depth-temperature curves tend to be straight lines, whereas, on the flanks, the curves tend to be convex, as shown by the diminution of the reciprocal gradients from 62.1 to 51.3 ft. per degrees Fahrenheit for the greater depths. The fact that the observed isogeotherms (Fig. 5) tend to spread apart on the flanks of the structure and the further fact that the depth-temperature curves tend to approach the same temperature

at a depth which is not very great are additional proof that most of the curves are convex toward the depth axis.

Local variations are found chiefly over salt domes and granitic plugs or ridges in which the height *cd* of the ridge, Fig. 5, above the general level of the basement rocks is several hundred feet or more. Such structures are frequently easy of detection because of the small values of $1/b$. For example, Homer, La., is probably a salt dome, and Teapot, Thermopolis, Salt Creek, Lance Creek and Lost Soldier in Wyoming are recent uplifts in which the basement rocks over small areas have been brought close to the surface of the ground. At Salt Creek, the granite is at a depth of about 1 mile beneath the surface at the top of the dome. In some structures, the rise of the isogeotherms over the folds may be imperceptible. In fact, Strong[58] states that limestone folds covered with saliferous strata cause depressed isogeotherms.

Regional variations depend largely on the depths to the basement rocks. Interesting evidence in support of this hypothesis has been found in southeastern South Dakota and eastern Nebraska,[59] central Oklahoma, and in the Appalachian trough extending from western New York through western Pennsylvania to the central part of West Virginia.[60] The temperatures increase as the depths to the basement rocks decrease, and in certain areas this relation implies that temperature gradients can be correlated empirically with gravity anomalies. That is, the intensity of gravity is highest in some of these areas in which the depths to the basement rocks are the least and the temperatures in the sedimentary strata are the highest.

TEMPERATURE GRADIENTS IN BASEMENT ROCKS

Two sets of reliable observations have been made in basement rocks—one by Drs. Krige and Pirow[61] in the Carnarvon district, South Africa, and the other by Dr. D. F. Hewett[62] in two wells in Georgia.

The Carnarvon well penetrated 2,603 ft. (793 m.) of sedimentary strata, represented by shale, limestone and sandstone, and 2,400 ft. (732 m.) of crystalline rocks, represented by granite and gneiss. The reciprocal gradient ($1/b$) in the sedimentaries is 1°F. in 44.2 ft. (1°C. in 24.2 m.) and, in the granite and gneiss, the value is 1°F. in 82.6 ft. (1°C. in 45.3 m.).

The Georgia wells are located near Griffin and La Grange. Observations were made at intervals of 100 ft. to a depth of 700 ft. in the Griffin well and 618 feet in the La Grange well. The reciprocal gradients are, respectively, 131.5 ft. per degree Fahrenheit (72.1 m.

per degree centigrade) and 124.7 ft. per degree Fahrenheit (68.4 meters per degree centigrade).

Dr. Hewett kindly prepared for the present writer the following very interesting note on the geological history of the area in which the wells are located:

Both of these wells lie within the Dadeville belt of metamorphic rocks of which the most widespread is that generally known as the Carolina gneiss. This rock is the oldest recognized unit of the belt all of which are pre-Cambrian in age. The areas in which the wells are drilled were probably deeply eroded in early Paleozoic time, covered with a thin layer of Paleozoic sediments which with an additional layer of crystalline rock was eroded in post-Paleozoic time. The present surface probably lies at least 10,000 feet and maybe 25,000 feet below the surface of the earth when the Carolina gneiss was formed.

The relative magnitudes of the reciprocal gradients in the Carnarvon well are doubtless due to the difference in the conductivities of the rocks. The small values of these quantities however, may be the result of subsidence, elevation and erosion in the order named; for according to theoretical deductions by Dr. Jeffreys[63] in 130 million years the temperature at the bottom of a sedimentary column that reaches a depth of 10 km. may rise to the extent of about 250°C. Once this condition exists, elevation and erosion would tend to keep the temperatures near the surface of the earth too high.

Metamorphism and subsidence were probably minor factors in the area in which the Georgia wells are located. The chief factors here seem to be the large-scale uplift and erosion that Dr. Hewett describes. The fact that hot rocks were brought from great depths to the surface of the earth indicates that the recorded reciprocal gradients, about 130 ft. per degree Fahrenheit (71.3 m. per degree centigrade) are again too small in comparison with a gradient in an undisturbed earth. That is, the gradients are dependent on factors other than lapse of time as determined from the uniform cooling of a homogeneous earth. On the other hand, the low temperatures in northern Ontario are probably the result chiefly of lapse of time. The only other important factor here seems to be the ascending waters which produced the metalliferous deposits. During the deposition of the metals and for a long time thereafter, the temperatures near the surface of the ground would be too high, but, after the lapse of millions of years, the removal of large quantities of heat by both convection and conduction would ultimately result in abnormally low gradients.

TEMPERATURE GRADIENTS IN LAVA BEDS

Deep wells drilled in the lava flows of Oregon and Washington afford an unusual opportunity to determine the temperatures[64] in these

areas of unusual geologic interest. Instead of smooth curves like the one shown in Fig. 4, we have curves made up of segments which intersect each other at sharp angles forming a series of steps as shown by the depth-temperature curve of the Burns well (Fig. 6). The flat segments

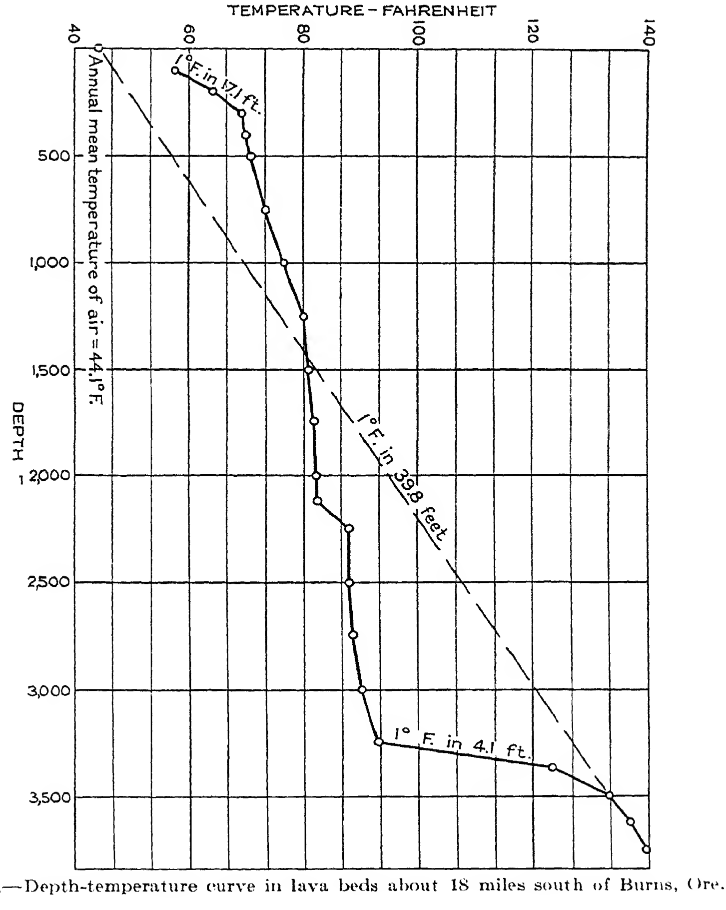


FIG. 6.—Depth-temperature curve in lava beds about 18 miles south of Burns, Ore.

probably represent convection of water (Fig. 6) in porous lava, and the steeply inclined segments represent the temperatures in strata of compact basalts.

The flows occurred before the latter part of the Pleistocene epoch, about 10 million years ago. The rock temperatures are unusually

high. Boiling-point temperatures are of frequent occurrence and at depths oftentimes of less than 1,000 ft.

A knowledge of the mechanism of a lava flow would enable us to understand the temperature gradients. Prof. Daly[65] believes that lava flows originate in the glassy substratum at a depth of about 40 miles. The lava is brought to the surface as the result of expanding gas and the release of rock pressure at the bottom of fissures that penetrate the earth's rocky shell. The cooling effect of the expanding gas would more than offset the heat developed by friction in the ascending lava. Chemical reactions would increase the temperature of the lava at the surface, but these effects are short-lived in comparison with the effects produced by bringing enormous quantities of hot lava to the surface of the ground. The removal by convection of such enormous quantities of heat from the areas immediately surrounding the fissures must ultimately result, in extremely long intervals of time, in reducing the temperatures much below that which would have been obtained by conduction alone. In the meantime, in the earlier stages of the cooling process, the flow of heat from a lava flow would approximate very roughly to the flow of heat from the earth immediately after solidification. In this sense, the large differences in the gradients in the old as compared with the recent lava flows may depend chiefly on lapse of time. Millions of years must elapse before this low temperature state is reached in Oregon and Washington.

SUMMARY

Our knowledge of the elevations of the isogeotherms in continental masses is very incomplete, for as a rule observations have been concentrated in oil fields and mining areas in which the geological conditions are highly abnormal. Observation leads us to expect, however, that lower temperatures, and probably very much lower temperatures, prevail in the large synclines surrounding the relatively small areas on which domes and anticlines are located; and, furthermore, because of the transfer of enormous quantities of hot water to the surface of the earth in metalliferous areas, it is to be expected that, with sufficient lapse of time since the waters reached the surface, the isogeotherms in these areas are depressed below their normal positions.

In areas of sedimentation, the isogeotherms tend to reflect the depths to the basement rocks. In central Oklahoma, for example, the isogeotherms immediately above the basement rocks tend to parallel the basement floor.[66] In general, then, the isogeotherms in sedimentary strata are usually elevated over domes and anticlines, whereas in the adjacent synclines they are depressed and their distribution in

the vertical is related to the depths to the basement rocks in the sense that the temperatures diminish as the depths to the basement rocks increase. In some areas, this relationship implies that high temperatures and positive gravity anomalies are related to each other.

In the course of geological ages, as a result of subsidence, elevation and erosion, the gradients have oscillated back and forth through a wide range of values so that an average gradient is more nearly a measure of anomalies than lapse of time as determined by a uniformly cooling globe. Only a rough estimate can be made of an average gradient in the sedimentary areas of a continent; but in the United States the reciprocal of the average gradient is almost certainly greater than 60 ft. per degree Fahrenheit (32.9 m. per degree centigrade) and possibly as large as 110 ft. per degree Fahrenheit (60.4 m. per degree centigrade).[66]

Undisturbed strata and age of rocks are prerequisites for greatly depressed isogeotherms, as exemplified by the sedimentary strata in the southern Appalachians and the Permian basin and, in areas of metalliferous deposits, by Grass Valley, northern Ontario, northern Michigan and Witwatersrand.

Tables 18 to 21 show that a rate of about 50 ft. per degree Fahrenheit (27.4 m. per degree centigrade) is found either at the surface or at a depth of 1 or 2 miles over a considerable portion of the sedimentary areas of the globe. In areas of old rocks, this high rate is probably the result of radioactivity and the combined effects of subsidence, elevation, and erosion. Not much is known about the trends of the depth-temperature curves in the low-temperature areas of metalliferous deposits; but Tables 20 and 21 show that these trends are apparently maintained from 2,000 ft. to the greatest depths at which temperatures are recorded. This result is consistent with the hypothesis that long intervals of time have elapsed since the lavas and hot waters were brought to the surface of the ground.

ACKNOWLEDGMENTS

The opinions of Messrs. W. D. Johnston, Jr., and J. B. Mertie, Jr., on processes of ore deposition were of great value to the present writer in his attempt to interpret the temperature gradients in areas of metalliferous deposition. Messrs. G. F. Loughlin, G. R. Mansfield, W. W. Rubey and C. H. Dane read the manuscript and offered helpful criticisms and suggestions. Mr. H. C. Spicer evaluated the thermal constants in Tables 18 to 21 by the method of least squares. To all of these contributors the writer gratefully acknowledges his indebtedness.

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CHAPTER VII

THE COOLING OF THE EARTH AND THE TEMPERATURE IN ITS INTERIOR

B. Gutenberg

(Professor of Geophysics, California Institute of Technology, Pasadena, Calif.)

The first period after the origin of the earth was characterized by relatively rapid changes. Mercier[24] has pointed out that any reasonable choice of characteristics of a gaseous globe having the mass of the earth and an initial temperature of 6000° absolute at the center leads to the conclusion that in a thick outer layer iron must have condensed immediately into liquid metal and consequently precipitated in drops toward the center. Jeffreys (Ref. 1, page 79) has estimated that the crust became solid within 15,000 years of its birth and soon afterward cooled at the surface sufficiently for an ocean to condense. An attempt to describe the changes during this time has been made by P. G. Nutting.[2] As V. M. Goldschmidt[3] has pointed out, the separation of the layers in the interior of the earth can be compared with the separation of ores in a blast furnace (see Chap. IX). Thus, during this time the core probably originated, surrounded by the mantle and a thin layer of still lighter material. Very probably, the differences in density between the various continental layers were large enough to prevent convection currents between one layer and another. However, in the Pacific area and in the deeper parts of the mantle, such currents possibly have existed and may still exist according to theoretical conclusions of H. Jeffreys (Ref. 1, page 140) and Pekeris.[4] As soon as the outer crust had solidified, the cooling proceeded very much more slowly. The details of these processes depend very much on the properties of the materials involved. For example, if in a layer the density of the crystalline form is noticeably higher than that of the glassy, crystalline blocks forming at the surface would sink down; otherwise they would stay at the top. If there is an increase in the melting point with depth greater than the increase in temperature in a given layer, its crystallization could begin at the bottom. The question of how long after the beginning of the crystallization the present "oldest" parts of the continents were formed, and similar problems, cannot be answered until we have more accurate data on the physical

constants and processes that are involved. For the same reasons, it is not yet possible to draw from the historical development of the earth conclusions as to the state of its interior.

The cooling of the crust of the earth after its crystallization has been studied by many [Refs. 1 (pages 138–160) 5, 6, 7]. The problem is much complicated by the fact that the heat escaping at the surface of the earth not only is due to the cooling but contains, besides, heat produced by the disintegration of radioactive matter. This, in turn, has changed in the course of time, depending on the disintegration constant. As Brewer[8] has pointed out, the disintegration constant of K^{40}, which is the radioactive isotope of potassium, leads to the conclusion that in the earlier history of the earth K^{40} was more abundant than it is now. The quantitative side of this and similar processes needs further investigation. On the other hand, geologists find that during certain periods, for example in the Archean, the earth was especially hot (see Chap. III).

The present change of the temperature T with the depth z, near the surface ($z = 0$), is given, according to Jeffreys (Ref. 1, page 150) with sufficient approximation by

$$kG_0 = k\left[\frac{\partial T}{\partial z}\right]_{z=0} = \int_0 P\,dz + kA, \tag{15}$$

or

Heat that escapes = heat from radioactive matters +
heat from cooling (including crystallization).

k is the thermal conductivity, P the rate of generation of heat due to radioactivity per unit volume and A represents the effect of the cooling of the earth, including crystallization. The value of A depends on various quantities within the earth, especially the gradient of the temperature in the mantle and the thermal conductivity k. As both are not known even approximately except for very moderate depth, the value of A is very doubtful. The heat arising from crystallization is probably less than the error in the calculation of the cooling in the crystalline part of the earth. H. Jeffreys (Ref. 1, page 151) has estimated that A is probably less than 4°C. per kilometer under the supposition that in the granitic layer $k = 0.006$ and in the deeper parts of the mantle $k = 0.004$, the specific heat is 0.20, the density is 3.3, the melting point of the deep-seated rocks under ordinary pressure is 1400°C. and its increase with the depth $m = 3$°C. per kilometer. The most serious assumption which he makes is that, in the deeper parts

of the mantle, the heat conductivity is less than in the granitic layer. If it is really larger there, A may be noticeably larger.

$\partial T/\partial z$ is the vertical gradient G of temperature. If we assume that the earth's crust consists of a number of layers with the thickness $D_1, D_2, D_3 \ldots$ and with the thermal conductivities $k_1, k_2 \ldots$ which are assumed to be constant in each layer, as well as the rates of generation of heat $P_1, P_2 \ldots$, we find from Eq. (15)

$$\frac{P_1 D_1}{k_1} + \frac{P_2 .}{k} + : \qquad + \text{ about } 4^\circ\text{C per kilometer.} \qquad (16)$$

The value of G_0 has to be adopted from the observations (see Chap. VI). Jeffreys has taken 32° per kilometer. A smaller value seems now to be more probable than a larger one. The sum of the terms $P\,D/k$ in Eq. (16), therefore, scarcely exceeds 30° per kilometer and is probably over 15° per kilometer. This means that the largest part of the heat current through the surface which is of the order of 10^{-6} cal. per second per square centimeter or of 5×10^{12} cal. per second for the whole earth is produced by radioactive matter within the earth. It should be emphasized that these figures may easily be inaccurate by a factor 2 or even more.

In order to find P, the amount of radioactive matter per gram of rock must be known, the heat generated by 1 g. of radioactive matter and the density of the rock, to compute the heat per unit volume. Data concerning the amount of radium and thorium in rocks have been collected and discussed recently by H. Jeffreys.[9] Some repre-

TABLE 22

RADIUM AND THORIUM CONTENT AND CORRESPONDING HEAT PRODUCTION IN ROCKS

	Content of		Heat produced, cal./g./sec., by					
	Radium, g. per g. rock $10^{-12}\times$	Thorium, g. per g. rock $10^{-5}\times$	Radium $10^{-14}\times$	Thorium $10^{-14}\times$	Sum $10^{-14}\times$	cal./cc. $10^{-13}\times$	k, $\frac{\text{cal.}}{\text{sec. cm. deg.}}$	P/k, deg./km./1 km. rock
Sediments	1.4	0.5	5.5	3.1	8.6	2¼	0.005	0.5
Granite *a*	1.6	0.8	6.2	4.9	11.1	3	0.006	0.5
Granite *b*	4.7	2.8	18.3	17.1	35.4	10	0.006	1.7
Granite *c*	4.4	3.3	17.2	20.1	37.3	10½	0.006	1.7
Basalt	1.0	0.9	3.9	5.5	9.4	2½	0.005	0.5
Plateau basalt	0.7	0.5	2.7	3.1	5.8	1¾	0.005	0.3
Peridotite	0.8	0.6	3.1	3.7	6.8	2	0.007	0.3
Eclogite, dunite	0.4	0.3	1.6	1.8	3.4	1	0.009	0.1

sentative values taken from his and other data are given in Table 22 together with the heat produced per gram of rock under the assumption that 1 g. radium emanates 0.039 cal. per second and 1 g. thorium 6.1×10^{-9} cal. per second. The granite *a* is an average for North America, Greenland, Iceland, Scotland, Ireland and Japan; *b* for Finland; *c* for the Alps. The amount of calories produced per cubic centimeter by radium plus thorium has been found upon the assumption that the density is 2.7 for the sediments, 2.8 for granite and basalt, 2.9 for plateau basalt and 3.2 for the others. Finally, the last column has been calculated on the assumption of the thermal conductivity k as given in the preceding column.

Unfortunately, the radium content as well as the thermal conductivity differs widely not only among various samples of the same type of rock, but probably very much with depth. The large differences in heat production have been studied by Jeffreys.[9] The data given for granite in Table 22 give an example.

Another complication arises from the fact that sometimes younger material has a higher content of radioactive elements than older material of the same type. Thus A. Holmes has found for Finland granites the following average contents (in g./g. material):

	Oldest	Younger	Youngest	
Radium	2.4	4.6	6.2	$\times 10^{-12}$
Thorium	8.7	26.7	58.5	$\times 10^{-6}$

though according to Joly [Phil. Mag., (6) **18**: 577 (1909)] the radium content of the lavas of Vesuvius increases from 2.8×10^{-12} for prehistoric lavas to 7.8×10^{-12} for lavas of 1631, to 13×10^{-12} for lavas of 1832 and to 16×10^{-12} g./g. material for lavas of 1906.

Laboratory measurements of the thermal conductivity k at a pressure of 1 atmosphere are given in Table 23. Its value under the conditions at even moderate depths is highly hypothetical. Thus far, two ways have been suggested to find its order of magnitude. From investigations of fluctuations in the magnetic field, Chapman[10] and collaborators concluded that a gradual increase in the electrical conductivity begins at a depth of less than 200 km., attaining at a depth of 700 km a value possibly one hundred thousand times as large as has been found for the rocks near the surface. McNish[11] has pointed out that there may be a similar increase in the thermal conductivity, if the Wiedemann-Franz relationship holds in this case. Wiedemann and Franz found from experiments in 1853[12] that for many materials

the thermal conductivity is proportional to the electric conductivity. Further experiments have shown that this is a good approximation for many metals and some other materials. However, it does not hold at very low temperatures, where the metals have electric "supercondue-

TABLE 23

THERMAL CONDUCTIVITY FOR VARIOUS TEMPERATURES AT A PRESSURE OF 1 ATMOSPHERE

(Unpublished values by Francis Birch and Harry Clark. Smoothed values, in cal./sec. cm. °C. in units of the third decimal place.)

t, °C.	0	50	100	150	200	300	400	500
Quartz:								
∥ axis	27	22	19	17	15	12	10*	
⊥ axis	16	13	12	11	9.7	8.4	7*	
Quartzitic sandstone:								
∥ bed	14*	12	11	9.7	9.0			
⊥ bed	13	11	10	9.4	8.7			
Granite:								
Rockport 1	.....	7.8	7.2	6.8	6.5	5.9		
Rockport 2	.....	8.3	7.7	7.2	6.8			
Westerly, R. I	5.8*	5.6	5.4	5.3	5.1			
Gneiss, Pelham:								
∥ foliation	.....	7.0	6.6					
⊥ foliation	.....	....	4.8					
Syenite, Ontario	.....	5.3	5.1	5.0	5.0			
Anorthosite:								
Transvaal	4.4*	4.5	4.5	4.6	4.7			
Quebec	4.1*	4.2	4.2	4.3	4.3	4.5		
Diabase:								
Maryland	.....	5.4	5.4	5.3	5.4			
Vinal Haven	5.2*	5.2	5.1	5.1	5.0	5.0		
Mt. Holyoke	5.0*	5.0	5.0	5.0	5.0	5.0	5.1	
Gabbro:								
French Creek	.....	5.4	5.3	5.2	5.1			
Wisconsin 1	.....	4.6	4.7	4.7	4.8			
Wisconsin 2	4.8*	4.8	4.8	4.8	4.8	4.8	4.8	
Pyroxenite, Transvaal	.. ..	9.2	8.5	8.1	7.8	7.3		
Dunite:								
North Carolina 1	.....	10.0	8.8	8.0	7.5			
North Carolina 2	.....	11.4	10.1	9.4	8.8			
North Carolina 3	.....	10.1	9.3	8.6	8.1			
Glass:								
Silica	3.3*	3.4	3.5	3.7	3.8	4.1	4.4	5.0
Pyrex	2.9*	3.0	3.2	3.3	3.5	3.7	4.0	4.4
Obsidian	3.2*	3.4	3.5	3.6	3.7	4.0	4.3	4.5
Diabase	2.7*	2.9	3.0	3.1	3.3	3.5		

* Extrapolated values.

tivity." No similar increase in thermal conductivity has been found. Thus, the conclusion that the increase in electric conductivity in the mantle of the earth is paralleled by a similar increase in thermal conductivity, may be open to question.

Another way to estimate changes in thermal conductivity has been pointed out by Bridgman.[13] From theoretical considerations he has found that the thermal conductivity is proportional to the velocity of longitudinal waves and inversely proportional to the square of the mean distance of separation of centers of molecules. Bridgman is of the opinion that this formula should apply approximately to conditions in the earth.[14] As the velocity of longitudinal waves increases with depth and the distance of the centers of molecules decreases, a noticeable increase of the thermal conductivity in the interior of the earth should be expected. All indications point to a considerable increase of k with increasing depth in the interior of the earth.

If we assume that the uppermost layers in the continents consist of a few kilometers of sediments, on an average about 15 km. of granite, underlaid by about 20 km. of basalt, the value of P/k in the first 40 km. of the continents is found to be about equal to the value found from G_0. This would mean that all the heat produced by radioactive matter originates in the uppermost 40 or 50 km. of the earth. However, we must consider the probable increase in k with depth. If we try to correct with this consideration in mind, we reach the conclusion that probably most of the heat produced in radioactive processes and escaping the surface of the earth originates in the uppermost part of the earth. The fact that apparently less heat comes out of the deeper layers is due to several reasons. (1) The heat conductivity probably increases noticeably with depth; for this reason, the ratio P/k decreases in a similar way. If there is a "superconductivity" under very high pressures, the corresponding deep layers would no longer contribute appreciably to the heat flowing through the surface of the earth. (2) The amount of radioactive matter per volume unit seems to decrease with depth, which means that P decreases. Goldschmidt and other geochemists have drawn this conclusion from theoretical reasoning. Table 22 indicates it for the materials occurring in the earth's crust.

The average radium content in stone meteorites is about one-tenth of that of eclogites and dunites, that in iron meteorites about one-hundredth,[15] in general between 10^{-14} and 5×10^{-14} g./g. meteorite.

The fact that the approximately known heat production in the upper 30 or 40 km. of the earth's crust seems to account for the heat current through the surface has been considered a major problem of

geophysics. Though many scientists believe that the decrease in radioactive material with depth and the increase in the conductivity are sufficient to explain this result, others are of the opinion that very much more heat is generated in the interior than escapes from the surface. One more unknown factor is the additional heat produced by radioactive sources not mentioned so far (potassium, rubidium).[16] Possible changes of radioactivity in time have been mentioned above.

There is a similar difficult problem in the field of the mechanics of the earth's crust: to explain the energies that produce mountains. There seems to be a possibility that a fraction of heat is transformed into kinetic energy, *e.g.*, used up by subcrustal currents due to differences in temperature; subcrustal flow of this type due to the differences between oceans and continents has been studied by Pekeris,[4] who found that stresses of the order of 10^8 dynes per square centimeter may be produced in this way. Investigations of such problems are handicapped by the fact that the conditions are frequently not even approximately known.

J. Joly[17] considered the data available as a proof that at present noticeably more heat is generated inside the earth than escapes. He tried to combine this assumption with the findings of geologists that, during the history of the earth, periods of extended activity (*e.g.*, during Carboniferous and Permian times and during the Tertiary Period) alternated with relatively quiet periods (*e.g.*, the Triassic, and Recent). His conclusions were that, during the quiet periods, heat accumulates in the earth's outer parts whereas during the "revolutions" more heat escapes, partly in larger volcanic processes, partly as a consequence of a melting of parts of the crystalline crust. The details of the theory have been disproved [Refs. 1 (page 323), 18, 19.]

The heat escaping now from volcanoes is only a small fraction of the heat passing through the crust by conduction. The heat current through the crust has been found to be of the order of 5×10^{12} cal. per second. The heat escaping in lavas has been estimated by F. Lotze[19] as at least 2×10^{10} cal. per second. Even if we assume that the oceans contain a few times as much lavas as the continents and add an amount for the heat escaping in volcanic gases and from intrusives, the total will remain of the order of a few per cent of the heat that is lost by conduction through the crust.

Though geophysicists in general favor a solution that supposes the cooling of the earth, most geochemists consider the possibility of at least a temporary increase in the temperature of the earth's crust, rather than admit a theory that allows an appreciable amount of radioactive matter only in the upper 40 km. of the earth. To decide

among the various hypotheses requires better data on the quantities involved. The most doubtful is the thermal conductivity k at the pressures that are involved. If it increases sufficiently with increasing pressure, the main difficulty of the problem would be removed, and a moderate content of radioactive matter of the order of that found in stony meteorites throughout the mantle of the earth would still not prevent its cooling.

If one supposes a law for the change of the content of radioactive matter with depth, the cooling since a given instant (*e.g.*, the time of solidification) can be calculated[20] under the assumption that no heat transfer by convection currents exists. The results indicate that even under the assumption of a relatively small content of radioactive matter the amount of cooling decreases very rapidly with depth unless k increases rapidly. As this possibility must be considered and as, furthermore, we know neither the distribution of radioactive matter nor the possible effect of subcrustal currents, the results, at the best, give only the order of the cooling, since the solidification of the crust about 100°C. at a depth of 500 km. and only a few degrees at a depth of 1,000 km. Except for the outermost parts of the mantle, the temperature in the interior of the earth is therefore possibly not far from the temperature at the time when the crust solidified.

As neither the temperature at that time nor the cooling since is known with sufficient accuracy, hypotheses as to the present temperatures differ very widely. In the earliest period of the earth's history, currents probably were active in reducing the temperature gradient in the interior, at least down to the core. Inside the core, another system of currents probably existed. For this reason it seems likely that, at the time when crystallization began at the surface, the temperatures in the interior did not differ greatly from, and were not far above, the melting point in the region below the level of crystallization. If materials have a "maximum melting point"[21] at a certain pressure (Table 24) and if for higher pressures the melting point decreases again, the temperature in the interior probably will be still more uniform. A high coefficient of conductivity in the deeper parts of the earth will act in the same direction.

For these reasons it seems probable that at the moment when crystallization began the temperature in the outer part of the mantle was given approximately by the melting point, which increases for basalt from about 1150°C. at the surface to about 1400°C. at a depth of 100 km., and more slowly beyond that point. For dunite the values seem to be about 1400° at the surface and between 1500 and 1600° at a depth of 100 km. After crystallization had taken place, cooling decreased the temperature at the surface to a value determined by the

TABLE 24

MELTING POINTS OF MINERALS AFTER F. VON WOLFF[6]

It is assumed that the melting point is given by $S_0 + ap + bp^2$, where p is the pressure in kilograms per square centimeter, or by $S_0 + mz + nz^2$, where z is the depth in kilometers. As b and n are negative, a maximum melting point results at the pressure $\frac{-a}{2b}$ and at the depth $\frac{-m}{2n}$. The latter is given in the last column. Q is the latent heat of fusion. The data need confirmation, and investigations are needed as to how far an extrapolation is possible from the limited range of pressures at which the observations were made.

Mineral	S_0	a	$-10^6 \times b$	m	$-n$	Q, cal.	$-m/2n$
Orthoclase	1,170	0.0029	0.205	0.8	0.017	3,560	23
Quartz	1,710	0.001	0.047	0.3	0.004	11,100	75
Diabase	1,150	0.005	0.038	1.5	0.003	5,600	500
Diopside	1,395	0.0065	0.041	1.9	0.0035	4,400	540
Basalt, augite	1,155	0.0096	0.040	2.8	0.0035	4,300	800
Olivine	1,250	0.005	0.020	1.5	0.002	5,600	750
Dunite	1,400	0.005	0.020	1.5	0.002		750

radiation of the sun. In the upper parts of the crust, the effects of the climate and of differences in the structure probably have produced noticeable differences in the temperature at given depths. Tentative calculations have been made by C. L. Pekeris.[4] He assumed that the mean temperature of the ground decreases in continental regions from the equator to the pole by about 60°C. (The temperature in the oceans below 2 km. is uniformly around 2°C.). Taking, furthermore, the mean depth of the Pacific to be 5 km. below the mean continental level and assuming the upper part of the crust under the Pacific to consist of 25 km. of basalt as against the corresponding part of continental crust made up of 10 km. of granite on top of 20 km. of basalt, he found that at the bottom of the layer which he considered the maximum temperature differences must be of the order of 200°C. Below, the differences decrease exponentially with depth.

The depth of the transition from the crystalline to the vitreous state may give us an important point of the curve giving the temperature as a function of depth. We have to suppose that this transition is caused by the fact that the temperature at this depth equals the melting point of the material. If we find indications for such a change everywhere at a certain depth, our assumption is very probably correct.

No definite answer has yet been given as to this depth of the boundary between the crystalline and the vitreous part of the earth; there may even be a number of crystalline layers at widely differing depths in the mantle, separated by vitreous layers. Not until the

melting curves are known under high pressures and the problem of the maximum melting point has been solved are further speculations possible on a theoretical basis. There is, however, some indication from seismic as well as from geological data that such a boundary exists at a depth of about 60 to 70 km.[22] In Fig. 7 the full-line curve indi-

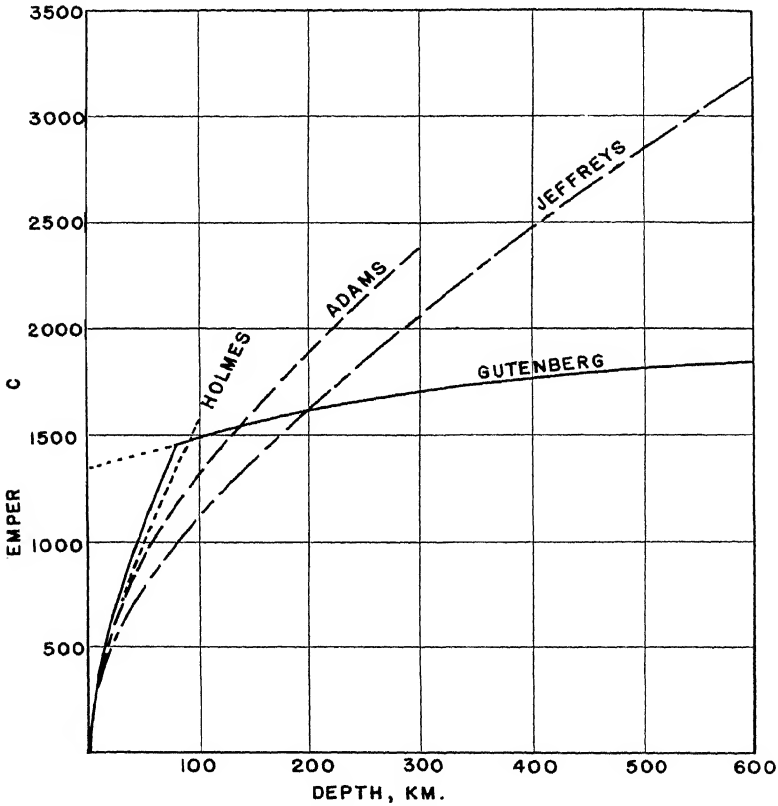

Fig. 7.—Temperature in the interior of the earth according to various authors.

cates the temperature that would follow by applying the considerations just mentioned. Its beginning agrees, in general, with the findings of F. von Wolff.[6] The continuation for depths below 70 km. is very tentative. It is based on the assumption that during the early history of the earth the viscosity of the interior permitted convection currents which produced a smaller temperature gradient than we have now in the crystalline crust, so that at the beginning of crystallization the temperature was represented by a curve of the type of the dotted line

beginning at 1400°C. in Fig. 7, but somewhat higher. Besides, as the coefficient of heat conductivity at greater depths is probably noticeably larger than that which we observe under conditions near the surface, a smaller temperature gradient is again to be expected in the deeper layers. Finally, the strong decrease, with depth, of the heat produced by radioactivity acts in the same direction.

Figure 7 contains, in addition, curves suggested by Jeffreys (Ref. 1, page 154), Adams and Holmes.[7]

The estimates for the temperature at the center of the earth, in general, are between 2000 and 4000°C. The extrapolation of the full-line curve in Fig. 7 would lead to a temperature of not much over 2000°. McNish[11] came to the conclusion that no satisfactory theory accounts for the earth's magnetic field unless the assumption is made that the temperature in the core is of the order of 2000°C. An argument which indicates that temperatures over 4000° are very unlikely and that even lower temperatures in the core are probable was made by E. Wiechert.[23] He contended that the assumption of higher temperatures would lead to too small a density for the core.

In summary it may be stated that probably the increase in temperature at greater depths is smaller than is generally believed; that the full curve as given in Fig. 7 represents a first approximation to the values; that the temperature in the center of the earth is probably closer to 2000 than to 3000°C.; that the interior of the earth is probably cooling very slowly; but that all conclusions may be changed considerably by new findings, especially concerning thermal conductivity under very high pressures.

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CHAPTER VIII

FORCES IN THE EARTH'S CRUST

B. Gutenberg

(Professor of Geophysics, California Institute of Technology, Pasadena, Calif.)

One of the major problems of geology has always been to account for the forces that produce mountains. Although numerous hypotheses have been published, this is still an unsolved problem, for each of the hypotheses has met with serious objections.

There are two groups of forces that affect the development of the earth's crust: the first includes forces that persist without great changes over long periods, such as those produced by cooling or by heating in the interior of the earth, those connected with its rotation or, finally, those depending on the fact that the earth's crust is not in hydrostatic equilibrium. The forces of the second group change relatively rapidly with time. Important examples are the forces produced by chemical processes, crystallization or melting. Erosion and sedimentation disturb the equilibrium of the crust. Although they persist over periods that are long in comparison with human history, each of them is a relatively short episode in the history of the earth. The same is true for the formation of icecaps during glacial periods and their melting afterward. Other examples in the second group are the forces developed by yearly changes of the vegetation or periodic changes in the distribution of air pressure over the two hemispheres of the earth. Finally, low pressure areas traveling over the surface of the earth and the tides have effects of still shorter duration.

From the beginning of research on the origin of mountains, the idea that they are an effect of the contraction of the earth due to its cooling has been discussed. The earth's body contracts; its outermost solid layer follows and is horizontally compressed. This compression produces stresses, which owing to the differences in elastic properties within the uppermost layers of the earth's crust, are not necessarily horizontal. As soon as these stresses surpass the breaking strength in a layer, the material ruptures; if the (viscous) strength is reached first, plastic flow sets in. In both cases, the lateral compression produces mountain chains. However, as has been pointed out in Chap. VII, serious doubts have been expressed by many whether the

earth is still cooling, for it is conceivable that more heat is produced by radioactive matter inside the earth than escapes through the surface. In any case, the output of radioactive heat is such as greatly to retard the cooling of the earth. Indeed, it now appears that the shortening of the radius of the earth during the second half of its geological history can have hardly exceeded 50 km. and probably is of the order of 10 km.[1] The shortening during recent geological eras is probably no more than 1 mm. per century, the corresponding compression of the crust being less than 1 square kilometer per century. All these estimates are based on the cooling of the earth and, therefore, are subject to the basic uncertainties mentioned in Chap. VII. The contraction theory will be dealt with more in detail in Chap. IX.

The cooling of the earth must produce strains. As there is now no cooling at the surface, and only a negligible amount in the very deep parts, there must be at some depth a level or surface of maximum cooling, which, at the same time, is a "level with no strain." Under certain assumptions Jeffreys (Ref. 2, pages 278–283) has calculated its depth to be of the order of 100 km. It is very difficult to allow for the additional contraction that must accompany any crystallization of deep-lying material, for this change of state, in general, involves decided shrinkage. Because of this and other uncertain terms in the premises, it cannot be said that computation can yet give even the approximate depth of the level of no strain.

A large number of forces have been investigated that are connected with the rotation of the earth. Those due to movements of the poles have been treated in Chap. XVI (pages 245–277) of vol. II ("The Figure of the Earth") of the National Research Council series on Physics of the Earth. These forces are very small. They have attracted attention mainly because the original theory of Spitaler[3] gave by far too large an order of magnitude for them. Correct equations have been given by Kravetz.[6]

The forces are produced in various ways.[4,5,6] If the earth's axis shifts, the equilibrium is disturbed and forces with a tendency to restore equilibrium originate. If the angular movement a of the axis is small as in the continuously observed movements, the theoretical vertical displacement z of a point in the geographical latitude φ and the longitude λ, measured from the meridian of the new position of the pole, is given to a first approximation by

$$z = \tfrac{1}{2}Ram(1 + k) \sin 2\varphi \cos \lambda \sin 1'',$$

where a = angular displacement in seconds of arc, m = the ratio of the centrifugal acceleration at the Equator to gravity there ($\frac{1}{288}$),

k = an elastic constant of the earth, probably slightly less than $\frac{1}{4}$, which can be derived from the movements of the poles (see Chap. XV), and R = radius of the earth (6,370 km.). In the case of the Chandler-Euler movements of the poles, the maximum distance of the pole from its average position is about 0.3″. With this value for a we find the maximum of z (for $\varphi = 45°$ in the meridian of the shift of the pole, $\lambda = 0°$) of 2 cm.; if the earth should acquire the new form of equilibrium, the maximum change of the terrestrial radius at a point on the surface from the center of the earth would be 2 cm. The strain produced in this way is therefore very small.

The acceleration connected with the maximum displacement z_m is approximately $4\pi^2 z_m/T^2$, where T, the period of rotation of the earth (sidereal day), is about 86,164 sec. With the data as given above, we find a maximum acceleration of a point at the surface of the earth to be only $1\frac{1}{2} \times 10^{-8}$ cm. per second per second.

Larger strains are found if we suppose the existence of extended movements of the earth's axis during geological eras. A displacement of the poles by even a few degrees requires maximum vertical movements exceeding 1 km. in order to restore the equilibrium figure of the earth. These movements, of course, do not produce mountains but have a tendency to shift the diameters of the earth into new positions. However, large strains would be connected with such an event, and, if great displacements of the poles occurred during the history of the earth, large strains would have been a consequence, especially in the meridian along which the poles were traveling.

If a continent shifts northward or southward over the deeper layers, the strain due to the change in latitude is much smaller than in the case of movements of the earth's axis. These strains depend on the thickness of the moving continent, on the differences in curvature of the earth's surface as a function of the geographic latitude and on the change of the ellipticity of the layers as a function of depth [see Chap. XIII, Eq. (65)].

While the forces discussed thus far are due to changes in the figure of the earth, other forces are produced by a shift of the axis of the earth in consequence of the change in the speed of rotation for a given point. The maximum acceleration b produced by a movement of the poles with a period T is given by

$$b = \frac{C}{\omega T} \sin \varphi, \qquad \text{where} \qquad C = \omega^2 aR. \tag{17}$$

ω is the angular velocity of the earth of about 0.7292×10^{-4} sec^{-1}. If we assume again a maximum of 0.3″ for a, we find $C_{\max}$ about

5×10^{-6} cm. per second per second. The period T of the movements of the poles is about 14 months. The corresponding maximum value of b is about 2×10^{-9} cm. per second per second. The strain produced in this way is, therefore, negligible.

Spitaler, finally, has discussed the changes in the centrifugal forces. Their maximum value is given by the value of C_{max}. Spitaler has pointed out that the horizontal stresses produced in this way may accumulate over larger areas in various latitudes and over large ranges in depth. In this way he finds a possible maximum pressure of the order of a few thousand dynes per square centimeter.

Observations[7] indicate the possibility that forces connected with the movements of the poles may act as trigger forces for earthquakes. Though it has been assumed generally that the movements of the poles produce stresses which act as trigger forces, Lambert[8] has pointed out that the stresses due to the diurnal tide in the earth body are so much larger that the changes in latitude cannot act as trigger forces and that another cause may produce both effects, movements of the poles and the corresponding periodicity of earthquakes. The sources of energy that maintain the "free" movements of the poles have been investigated and summarized by Jeffreys (Ref. 2, pages 239–244). He finds that seasonal variations in the distribution of air over the surface, the changes in the load of snow from one season to the other and periodical changes in vegetation may account for the energy. On the other hand, Lambert points to the possibility that expansions and contractions of the earth may produce both phenomena, the periodicity in earthquakes and the variation in latitude. He bases this view on Brown's hypothesis[9] that apparently irregular fluctuations in the motion of the moon involve an expansion or contraction of the earth over irregular periods of years, decades or even centuries. "The corrections to the moon's tabular place, as determined by observation, seem to vary most rapidly, at just about the times when there is a sudden change in amplitude or phase of the free motion of the pole, or a marked progressive shifting of the mean pole" (Ref. 8, page 136). The loss of energy due to internal friction seems to be small (see Chap. XV).

The stresses produced by changes in the speed of the earth's rotation have been investigated. If $\Delta\omega$ is the change in the angular speed of rotation of the earth (radius R) per second, the acceleration of a point in the latitude φ is given by $\Delta\omega R \cos \varphi$. This would give a maximum of about 1.3×10^{-3} cm. per second per second or about 1 milligal, if the period of the earth's rotation changes 1 sec. per year. The corresponding change in the centrifugal force is about 2×10^{-6}

$\cos \varphi$ cm. per second per second. A change in the speed of rotation of the earth would, of course, tend to change the figure of the earth.

One of the forces discussed most frequently is the *Polfluchtkraft.*[10] If we suppose that the lighter continents are "swimming" in denser material, the center of gravity A of a given continent will be at a slightly higher level than the center of gravity B of the denser material which would be at its place. The pressure of the continent downward and the pressure of the deeper material upward are equal and perpendicular to the respective surfaces of A and B. If these two surfaces were parallel, the forces would be in equilibrium. However, the surfaces converge toward the poles, owing to the decrease in flattening of the surfaces with depth, and, therefore, the forces acting at A and B are not in the same direction. Their resultant force is directed toward the equator and tends to shift the continents in this direction, hence the theoretical idea of a force competent to make the continents "flee from the poles"—a *Polflucht.*

The amount of *Polfluchtkraft* depends very much on the distribution of density with depth. Even the order of magnitude differs under various assumptions which are within the possible limits. To a first approximation, the force acting on a continent with the mass m, the height d above the surface of the material in which it swims and boundaries along the parallels a and b is given by

$$0.4\, m\, d\omega^2 (\cos^3 b - \cos^3 a),$$

where $\omega = 2\pi/86{,}164 = 0.7292 \times 10^{-4}$ sec.$^{-1}$ = angular velocity of the earth. The maximum values found in this way for a continent extending from a pole to the equator are of the order of 10^6 dynes per square centimeter, corresponding about to the pressure of a column of rock a few meters in height.

In a recent paper A. Prey[10] has considered the possibility that the continent undergoes not only a translation in the north-south direction but also a turning about an axis in the east-west direction. In a special case he found that the continent must rotate about this axis by $\frac{1}{2}''$ and must be translated toward the pole by 20 m. to restore equilibrium but that probably the forces are too small to produce any movement.

In spite of the fact, admitted by all, that the *Polfluchtkraft* is small, it has been considered of great importance in many geotectonic theories (see Chap. IX). There is the question, moreover, whether a swimming of the continents can be assumed.

Another force that has played an important role in the theory is supposed to shift the continents toward the west. It is known that the

tides rolling over the surface of the earth mainly from east to west produce friction, which has been discussed by W. D. Lambert in Chap. VI of vol. II of this series. However, the observations show that this effect is very small and that the theoretical "west-drift force" is negligible. Jeffreys (Ref. 2, page 322) has calculated that a west-drift force large enough to shift America westward by 50° in 3×10^7 years would stop the earth's rotation in a time of the order of 1 year.

Isostasy shows that at a certain depth, probably between 50 and 100 km., there is approximate hydrostatic equilibrium. However, at smaller depths there must be considerable stresses with the tendency to spread the higher parts of the earth's crust and especially the continental layers over the sima. The problem of such stresses produced by the inequalities in the upper layers has been investigated by Love.[11] He found that the maximum stress difference for a harmonic inequality of the first degree is found at the places where the gradient is steepest, not at the places of greatest elevation and depression. Under certain assumptions, it is found at a depth equal to one-third of the thickness of the layer of compensation. For harmonics of higher degrees, it moves toward the places of greatest elevation and depression. The amount of this maximum stress difference depends on the conditions. For an isolated mountain, 2 km. in height, Love finds the maximum stress difference to be about one-half of the weight of a column of this mountain. For a series of parallel mountain chains with a height of 4 km. of the crests above the valley bottoms, the maximum stress difference is about one-fourth of the weight of a column and is found beneath the crests. Thus, this force in places must be of the order of magnitude of 10^9 dynes per square centimeter. Strength prevents high parts of the earth's crust from flowing over the lower. There are, however, some indications that the continental layers are spreading slowly over the Pacific basin, possibly due to the forces just mentioned (see Chap. XI).

All the forces mentioned so far seem to be insufficient to produce the formation of the known mountains. The contraction theory possibly can explain a part of it, but a large source of energy is needed to explain the powerful changes in some of the geological epochs, *e.g.*, during the Tertiary. Chemical changes and radioactive processes (Chap. VII) can possibly produce such energies, but even their order of magnitude cannot be estimated. Various types of research seem to indicate that the source of large changes in the earth's crust is at considerable depths and that subcrustal movements precede and start the movements in the outer parts of the crust. Small subcrustal movements may be

started by the differences in temperature and thermal conductivity as between ocean bottoms and continents.[12]

All other forces of local type are small. If a low-pressure area moves across a continent, it produces an upwarp of the solid crust of the same wave form. The amplitude (from the rest line) is given by $ghb/2\pi\mu$, where g = gravity, h the deviation of the air pressure in centimeters of water (or in centimeters of mercury $\times$ 13.6), b = diameter of the low-pressure area, and μ = rigidity of the upper layers of the earth.[13] If we suppose, for example, a difference in pressure of 50 mm. mercury between the center of a high- and a following low-pressure area, distant 4,800 km. from each other, and a coefficient of rigidity of the order of 10^{11} dynes per square centimeter, the change in air pressure is accompanied by an elastic rise of the earth's crust of the order of 10 cm. The maximum tilt is given by

$$gh \operatorname{cosec} 1''/2\mu = 10^8 h/\mu \text{ seconds of arc.}$$

In our example, we find about 0.05″. The tilt actually observed during the passage of low-pressure areas has been used to calculate the rigidity in the earth's crust (see Chap. XIV).

The movements of the surface of the ocean are much larger, in our example about $\frac{1}{2}$ m.

A small force in an east-west direction is connected with vertical movements of crustal blocks due to the change in rotational speed.[14] This force is about 2,000 dynes per square centimeter, if a block of 1 square kilometer moves up or down by 1 cm. per year.

Small strains are connected with the changes in temperature in the earth's crust. If we suppose a daily range of temperature at the surface of 20°C., about 8° at a depth of 20 cm., and 2° at 50 cm., the total change in length is of the order of 0.005 cm. In a similar way, we find the change in length of a column of the order of 0.03 cm. during a year, supposing a range of the average temperature at the surface of 18°C., of 6° at 2 m., and of 1° at a depth of 6 m. The results depend much on the type of the ground, and the ranges differ widely.

The effect of changes in temperature has been observed in mines. Records of pendulums taken at depths down to 200 m. to find the tides of the earth's body showed clearly fluctuations in tilt due to the diurnal changes in temperature.

Long-period seismographs record a specific type of microseisms during times when the temperature drops below the freezing point. They are quite irregular waves with periods of $\frac{1}{2}$ min. or more; their amplitudes grow with increasing frost and may amount to a large

fraction of 1 mm.[15] In a region with wet sand the thickness of the freezing layer increases by about 3 per cent, depending on the pores.

Near the coasts, the tides produce movements due to the change in the load of the water. They have been recorded close to the source as well as at larger distances. The records of the tides of the solid earth are disturbed by such effects, which have their maximum in the direction of the largest effective ocean tides from the recording stations. As the periods of both types of tides are similar, it is difficult to separate them, and consequently the records of both effects originally had been misinterpreted as due only to body tides. Thus they seemed to indicate that the rigidity of the earth's crust is a function of azimuth. Even as much as 1,000 km. from the coast, tilts of the order of 0.001″ are produced by the change in load of water from low to high tide at the coast.

The tides of the earth's body, of course, produce strains. The maximum range of the tide with a period of ½ day is about 30 cm., the maximum tilt produced by it about 0.02″ and the maximum change in gravity of the order of 0.1 milligal.[16] The maximum tilt observed by means of pendulums and including tides, effects of changes in air pressure, load by rain and other changes is usually less than 0.1″, the maximum change in gravity due to the same sources, in general, less than 1 milligal.

Lambert[8] has pointed out that stresses due to the average maximum diurnal earth tide, which occurs about once a fortnight, are about twelve times as large as those due to the largest usual maximum displacement of the pole, so that the trigger effect of the variation of latitude with its much longer period may be disregarded. Correlations between the body tides of the earth and the occurrence of earthquakes seem to exist;[17] however, the results need further confirmation.

An intermediate type between the forces acting through geological epochs and those acting for a relatively short time are those which gradually change the load of a larger area during periods of between thousands and millions of years. Sources for such changes are sedimentation, erosion, forming of glaciers and icecaps on extended regions and melting of large amounts of ice. All these processes are very slow, and in most cases plastic flow at large depths prevents large gravity anomalies. In the case of sedimentation, the density of the material that is removed usually will be only slightly less than the density of the material that replaces it at depth in the zone of plastic flow, and, therefore, the new surface will be only slightly lower in altitude when the isostatic equilibrium has been restored. In the cases of sedimentation on land and changes due to forming or melting

of ice, the change in level is given by $h\left(1 - \frac{a}{b}\right)$, where h is the thickness of the ice, a is the density of the ice or sediments and b the density of the material at depth (about 3). When equilibrium exists again, the new surface will be at a higher altitude if material is deposited or ice forms, and at lower altitude than the surface of the ice when ice is melting. The changes connected with the forming or melting of icecaps are very complicated, as has been pointed out by Daly.[18] The displacements that arise from such changes in load have been investigated by N. A. Haskell,[19] under the assumption that the earth's crust has no strength even in the upper part and that it is incompressible.

If an icecap forms with the thickness h, the rock surface should finally be about $h/3$ lower than it was originally; in regions where the

Cause of force	Direction of force in the continents	Maximum of force or effect at present
Cooling of the earth	Compression	Decrease in radius of the earth of the order of 1 mm./century
Difference in level continents—oceans (Polflucht)	Toward equator	10^6 dynes/sq. cm.
Tidal and other friction	Westward	Very small
Secular movements of the earth's axis (?)	Vertical	Changes in radii of the earth by several km.
Chandler movement of poles (changes in latitude)	Horizontal	10^3 dynes/sq. cm.
Deviation from hydrostatic equilibrium of crust	Horizontal	10^9 dynes/sq. cm.
Chemical processes	Subcrustal currents	Probably large
Cosmic sources	?	?
Thermal differences between ocean bottoms and continents	Subcrustal currents	Several m./year
Changes in air pressure	Tilt	0.01″
Body tides	Tilt and vertical movements	0.1″ 30 cm.
Erosion, sedimentation (thickness d).	Vertical movements, subcrustal currents	$\frac{1}{3}d$ at surface
Forming or melting of ice	Vertical movements, subcrustal currents	$-\frac{1}{3}d$ (rock surface) $+\frac{2}{3}d$ (ice surface)
Variation in sea level (ocean tides, storms)	Tilt	1″
Freezing of ground	Vertical movements	1 cm.
Seasonal changes of temperature in ground	Vertical movements	0.1 cm.
Vertical movements of blocks in the earth's crust	Horizontal	10^4 dynes/sq. cm.

surface had been depressed by a load of an ice sheet with the thickness of d, it must move upward by about $d/3$ after the melting of the ice, to restore approximately isostatic equilibrium. The order of time that is needed for this process depends on the viscosity in the zone of plastic flow. In case of fast melting, it is probably between 10,000 and 100,000 years for a large glaciated area, and more for a river delta, depending on the area of sedimentation.

If the sedimentation occurs in water, the density of water must be subtracted from the density of the sediments. As the isostatic equilibrium seems to continue during sedimentation with sufficient approximation (see Chap. XII), the top of the sediments should not rise by much more than one-half of the thickness of the sediments.

The tabulation on page 173 gives a summary of the forces and their magnitude.

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CHAPTER IX

HYPOTHESES ON THE DEVELOPMENT OF THE EARTH'S CRUST AND THEIR IMPLICATIONS

B. GUTENBERG

(*Professor of Geophysics, California Institute of Technology, Pasadena, Calif.*)

One of the major problems of geophysics and geochemistry is to trace the development of the earth's crust after its solidification. The forces acting in the crust have been discussed in the preceding chapter. Their application to an early state of the crust should lead, theoretically, to the events that have changed the crust and to its present structure and condition. The normal procedure, however, is to look for theories that explain the present conditions. The difficulty with the logical method, starting with the beginning and coming down to the present state, is that neither the initial conditions nor the forces which have produced the changes are well enough known. The disadvantages of the second method, proceeding backward from the present conditions, are that the present situation is assumed to be characteristic for the whole geological past. Too easily, hypotheses are constructed which do not allow properly for all the forces and rhythms involved. Too frequently the assumption has been made that one single force or one simple process explains the whole history of the earth's crust.

The books and papers dealing with hypotheses on the development of the earth's crust are as the sands of the sea. As the processes involved may contain clues as to the interior of the earth, at least a few of the fundamental ideas are to be here presented with some fullness. Somewhat more detailed critical descriptions have been published by Nölke[1] and Gutenberg.[2] Many authors, while pointing out objections to hypotheses other than their own, have supplied useful bibliographies bearing on the subject.

OBSERVATIONS AND "LAWS"

A multitude of definite, observed facts are in hand, and these must necessarily control ideas about tectonic processes. No hypotheses can be considered that clearly contradict such facts. Other observations have been combined, and certain "laws" have been derived from them. In many instances these laws have been found by extrapolation.

The word *law* was first used in the preceding paragraph with quotation marks to indicate that these "rules" are not laws of nature—which we do not know—but have been derived from observations and may be "recalled" any time by scientists following the finding of new data or even a rediscussion of older ones.

The facts, forces, laws and hypotheses can be grouped in a few classes. The first has to do with the early history of the earth; the second, with subcrustal flow; the third, with contraction (or expansion) of the earth during its history; the fourth, with movements of the poles, the speed of rotation of the earth and changes in climate; a last group, with movements of the continents or large parts of them relative to each other.

A number of facts are given in Chaps. III, VII and XII. Others will be mentioned in connection with hypotheses with which they have been respectively considered. A number of forces, conceived of as of possible importance, are discussed in Chap. VIII.

The laws may be divided into two groups: those derived from applications of physics to the problems and those involving geological data mainly. Many results of the first type refer to plastic flow. The first three of the following list are based on findings of Geszti:[3]

1. The volume of a given body can be taken as constant during plastic flow.
2. Plastic flow requires stress differences larger than strength between at least two different directions. In connection with this law, it must be pointed out, however, that "creep" may occur under very much smaller stress differences. Gradual movements in the uppermost parts of the earth's crust, such as have been found by leveling in many regions or by the closing in of mines, are examples.
3. Plastic flow occurs in such a way that the energy needed to change the form is a minimum, considering the possible changes.

Geszti has considered, besides, a number of special examples.

The relationship between plastic flow and rupture may be expressed as follows:

4. If in a body that is undergoing plastic flow the stress difference in two directions increases faster, due to the acting forces, than it decreases from plastic flow, rupture will occur when the breaking strength is surpassed. This law has been used by Gutenberg in explaining the possibility that earthquakes

occur at a depth where plastic flow under very small stress differences is required to explain isostasy (see Chap. XI). Recognized long ago, this principle has been verified by the experiments of D. T. Griggs.[4]

The following is the law of isostasy:

5. Each vertical column of the earth's crust in regions that have not been disturbed recently, with a given radius (at least 10 km.) and extending down to a given depth (at least 60 km.), has approximately the same mass whether its column occurs in a continental, mountainous or oceanic region. However, neither the theory of Pratt (the density varies inversely as the height of the columns) nor the theory of Airy (the density of a given layer is constant, but under mountains the lighter surface material extends to larger depths) corresponds to the whole reality. Both explanations of the earth's relief contain truth: but in one region the first theory may be a better approximation; in another region the second theory is preferable.

A number of problems were investigated by H. Jeffreys,[5] especially in connection with the contraction theory. The following are two of his conclusions:

6. In any problem of mountain formation the crust can transmit the stresses perfectly for any distance, and failure takes place where the stress difference first reaches the strength of rocks (*op. cit.*, p. 288).
7. If the earth contracts by cooling, the vertical stress is practically the same as in hydrostatic equilibrium; the type of yield is in this case a crushing due to the horizontal pressure's being greater than the vertical pressure. Both are compressive stresses (*op. cit.*, p. 291).

A series of laws were derived by Stille[6] from geological results. The following is an abstract of them:

8. All mountain making has occurred during a relatively few phases, each of short duration. (This does not include orogenic movements.)
9. Tectonic activity occurs more or less simultaneously in different regions during each of those phases.
10. Any structural type may originate during any period.

11. There is always movement upward relative to the surface of the oceans during high tectonic activity.
12. Folding and block movements are due to the same forces.
13. The result depends mainly on the material and its mobility.
14. Changes during epeirogenic movements are very frequently in the same direction in different regions.
15. The main causes of epeirogenic movements are not local.
16. Orogenic and epeirogenic movements differ in the magnitude, not the nature of their cause. (There is some doubt about this law.)
17. The tectonic processes are a function of (*a*) the intensity of the forces (effect of pressure), (*b*) the strength of the material involved, (*c*) structural conditions at the beginning. The ratio of the first two is of greatest importance. Orogeny is the form of reaction to strong episodic pressure, epeirogeny the reaction to smaller secular pressure. A small pressure produces in a weak material the same effect as a greater pressure in stronger material.
18. Geotectonic processes occur in such a way that the energy needed is a minimum.

W. H. Bucher[7] has condensed the results of investigations on geological problems into about 50 laws. Some of them express ideas similar to those found by Stille. We mention the following more general laws, partly in the form of abstracts of Bucher's laws:

19. In the progress of crustal deformations, the direction of radial displacement is reversible.
20. On the present face of the earth, excessive heights and excessive depths of crustal deformation are of distinctly linear outline.
21. Long lines of depressions (*furrows*) and of elevations (*welts*) are closely associated and lie side by side in relatively long and narrow belts.
22. Laws 19, 20, 21 have been valid throughout the geological past, as far as can be judged from available records.
23. The frequency curve of elevations on the earth shows two pronounced maxima, corresponding to the ocean floors and to the continental platforms.
24. The typical orogenic cycle begins with geosynclinal depression and ends with a major uplift. Quiet sinking prevails in the first phase; the second consists of crustal folding separated by diminishing epochs of renewed geosynclinal sinking.

25. The structure of welts proves the existence of differential movements in the crystalline materials which approach the nature of plastic flow. The perfection of this plastic behavior increases with depth.
26. New mobile belts are found to resemble the next older (largely inactive) belts in trend and general pattern and to lie adjacent to or superimposed on them.
27. In a large way, the major movements of strand lines have affected all continents in the same sense at the same time.
28. The great transgressions of the sea upon the continents occurred in the intervals between the larger orogenic episodes.

Laws referring merely to the mechanism in the outermost part of the crust have been omitted; many of them can be found in the discussion of surface phenomena in Chap. III.

HYPOTHESES CONCERNING THE EARLY HISTORY OF THE EARTH

Hypotheses concerning the origin of the earth have been discussed by H. Jeffreys in Chap. II. There, as well as in Chap. III, the hypothesis has been mentioned that the moon may have originated from the earth. The theory indicates that this must have happened—if at all—practically at the time of the origin of the earth. Nevertheless, it possibly could explain the fact that the uppermost part of the crustal material is lacking in the bottom of the Pacific ocean. This is not necessarily to be explained by the assumption that the missing material has formed the moon; the density of the moon could be accounted for under this hypothesis,[8] as well as under the assumption[9] that the moon formed separately, but close to the earth. In the latter case, the Pacific basin may have lost its upper layers during the catastrophic lunar tides of the earth's crust at the time when the moon was very close to the earth.

Great changes must have occurred in the interior of the earth during its first stage. Possible chemical and physical changes that occurred on a large scale during this time and are still going on slowly have been studied by V. M. Goldschmidt.[10,11,12,13] In the gravity field of the earth, materials must have tended early to separate according to their densities. As in the case of pig iron, matte and slag in the blast furnace, there must have been, according to Goldschmidt, a separation of the terrestrial stuff—a molten metallic core, an overlying sulphide oxide melt, reaching up to a level about 1,000 km. from the surface (see Chap. X) and a silicate melt which is the remaining upper

part of the mantle. There is, besides, a gaseous phase, the atmosphere. Metals that are reduced more easily than iron went at once into the core; others, with a strong affinity to sulphur, went into the lower part of the mantle.

As crystallization went on, the outer part of the mantle separated into various layers. During the last stage of this change, which still continues and possibly is responsible for major tectonic processes, gases and solutions produced a large variety of changes in the crystalline crust. The relatively small amount of noble metals in the outer part of the crust is due not only to the early separation into the three layers but also to processes of the type just mentioned during relatively recent times. In his publications (*op. cit.*) Goldschmidt has discussed the conditions in detail for many elements. Of interest for the thermal history of the earth is his statement that thorium and uranium are to be expected preferably in the outer parts of the silicate layers.[10]

Goldschmidt, finally, has divided the elements into four classes: siderophile elements, concentrated in nickel iron (core of the earth); chalcophile elements, concentrated in sulphide melts (deeper part of the mantle); lithophile elements, concentrated in silicate melts (outer part of the mantle); atmophile elements, concentrated in the atmosphere.

The concentration of the elements into one or the other of these classes depends much on their atomic properties. Certain elements have so closely the same properties that they scarcely can be separated. Hafnium was discovered relatively late, although it forms about 0.003 per cent of the earth, owing to the fact that it has properties very similar to those of zirconium; it seems to hide behind it.

The atomic volume of the elements has a clear correlation with their occurrence in one class or another. If the atomic volume of the elements is plotted against the atomic number, the resulting curve shows maxima and minima. All siderophile elements (C,P,Fe,Co,Ni, Mo,Ru,Rh,Os,Ir,Pt) are near the minima of the curve; they all have small atomic volume; the chalcophile elements are on the sections where the atomic volume increases with the atomic number; they are followed by the atmophile elements, while the lithophile elements are near the maximum and on the decreasing sections of the curve. Goldschmidt has studied many of the problems connected with these facts.

Another fact of interest is the finding of Harkins[14] that in the earth as well as in meteorites each element with an even atomic number is, as a rule, more frequent than the preceding and the following element with an odd atomic number. The latter are less stable. Experi-

mental and theoretical investigations of Goldschmidt (Ref. 11, 1925, No. 5) have added further data to Harkins' results.

Hypotheses as to the composition of the whole earth may be based on geochemical reasoning and on comparison with other heavenly bodies, especially meteorites (Chap. V). There is a general agreement that iron takes the first place, the estimates being that it forms between 36 and 50 per cent of the material of the whole earth. Oxygen is second in amount, with an estimated percentage of between 22 and 30 per cent. Silicon follows with about 12 per cent, magnesium with 8 per cent, Ni and Ca with about 3 per cent each, Al with perhaps 1 or 2 per cent, S, Na and Ti with probably between ½ and 1 per cent and all others with less than ½ per cent.

The composition of the earth is influenced by the fact that the light gases escaped during the early history when the earth was very hot. Very probably, hydrogen and helium are still escaping.[16] No lines of these gases have yet been observed in spectra of luminous phenomena in the stratosphere, nor are there other indications of their existence there. The temperature in the upper stratosphere is possibly high enough to favor the escape of both gases in the course of time. As a consequence, light gases are much less abundant in the earth than in the sun. If one omits the light elements, it is found[17] that the percentages of the remaining elements, relative to that of iron, are very similar for the earth and for the sun. Besides the composition of the earth apparently corresponds roughly to 55 parts of stone meteorites plus 50 parts of iron meteorites plus 5 parts of troilite.[17]

Though Goldschmidt tried to picture in a general way how layering in the earth originated, Geszti[18] has investigated this problem more in detail. He tried to show that at the time when the crystalline crust formed the layers had not yet attained their present radii. As, in the deeper parts of the earth, gravity decreases with depth reaching zero in the center, all processes that are controlled by gravity must decrease in speed toward the center of the earth. Relatively small accumulations of materials that are heavier than their neighborhood will not sink toward the center of the earth. Only after larger masses of heavier material have accumulated may extended mass movements occur, which will produce, in turn, great disturbances at the surface. Geszti correlates such revolutions in the earth with the geological epochs in which large changes have occurred. In several other papers[19] he has tried to explain the origin of oceans and continents. Here he pointed out that, when condensation occurred during the early history of the earth, the originally deeper parts of the earth must have experienced higher pressures than the higher parts and that the existing

differences have been increased. The details are produced by physical-chemical properties of the silicate melts.

HYPOTHESES BASED ON SUBCRUSTAL CURRENTS

As has been indicated in the preceding section, there must have been large subcrustal currents in the young earth. Many writers have expressed the belief that the strength in the interior of the earth prevents any currents today. The results of geophysical research during the most recent decades, however, leave no doubt that such currents still exist, although of relatively low velocity. Isostasy would be impossible without such currents. If, by sedimentation or erosion or forming of icecaps, material is removed from the surface or added to it, equilibrium can be restored only by subcrustal flow. In these examples, the subcrustal movements are a consequence of changes produced by disturbances at the surface.

There are, however, a rather large number of theories in which the subcrustal flow is considered the primary cause of movements at the surface. Mainly of historic interest are the "plutonic" theories of L. v. Buch and Alexander v. Humboldt. Some more recent theories in which movements of partly plutonic origin have been considered, such as those of Deeke,[20] Rothpletz[21] and W. Penck,[22] involve the assumption of subcrustal movements without direct mention of it. O. Ampferer[23] was the first to bring to full emphasis the hypothesis that subcrustal currents have been the chief cause of mountain making and other changes at the surface of the earth, but he did not explain the cause of the subcrustal currents.

Many hypotheses have since been developed which either are based on such currents or implicitly assume them. The main theoretical difficulty lies in the fact that thus far no simple theory has been developed which accounts to a good approximation for plastic processes (including creep). It is overlooked by many that neither Hooke's law nor the theories of strength, viscosity and internal friction have been derived from mathematical-physical theories of matter, but are based on observations. Thus, if we find a contradiction between the interpretation of observations and results derived from the theory of viscosity, this interpretation is not necessarily wrong. For details, reference is made to Chap. XV.

There are probably various types of primary causes of subcrustal currents. K. Andrée[24] considered mainly changes in volume during crystallization. Apparently, the effects produced in this way are by far too small. The same is true for the effects of differences in differentiation which have been suggested by C. Mordziol.[25] A

combination of various processes, such as chemical and physical changes, crystallization and subcrustal currents especially, lead to the result, according to E. Kraus,[26] that the continents periodically increase in size by mountain ranges that have been added to them after differentiation from the suboceanic material.

Subcrustal currents play an important role in the hypothesis of Joly which has been mentioned in connection with the cooling of the earth (Chap. VII). According to his hypothesis, more heat is produced in the interior of the earth by radioactive material than escapes through the surface. Drastic revolutions in the earth's crust from time to time permit the escape of this heat; however, thus far no mechanism has been described which explains the periodicity of this process. This hypothesis has been modified by A. Holmes,[54] G. Kirsch[100] and others.

Movements in the interior of the earth of a cyclonic type have been suggested, for example, by R. Schwinner,[27] S. Fujiwhara,[28,29] L. Du Toit[88] and F. Rinne.[30] S. Fujiwhara has referred to structural evidences showing spiral forms at the surface of the earth corresponding to the flow of air in a cyclone. The type of flow that has been sketched by Schwinner has been assumed by Vening Meinesz.[31] He started from the fact that in regions of present tectonic activity there is frequently a belt of considerable negative gravity anomalies, accompanied usually by two more or less clear zones of positive anomalies, one at each side (see Figs. 21 and 22). He points out "that there are two possible ways of explaining an excess of gravity over such an extensive area, lateral compression in the earth's crust or descending currents in the substratum, and as this latter assumption likewise involves lateral compression of the crust, both explanations concur in corroborating the above result." Descending currents "must be fed by converging horizontal currents below the crust and these currents must exert a viscous drag on the crust which will lead to compression." In the region of the negative anomalies, a great downward protuberance is developing much faster than the upward protuberances. Vening Meinesz has tried to describe the process more in detail The drag of the currents causes the crust to buckle, the surface layer crumples up and ridges form. Temperature effects produce a rising of the tectonic area at a later stage.

From a gravity profile across Java and the region of the negative anomalies to the southwest, Vening Meinesz has concluded that the crustal shortening of a crust with a thickness of 25 km. would be of the order of 50 km. corresponding to a cross section of the root of the order of 1,200 sq. km.

No tectonic hypothesis can be considered that does not agree with the observed gravity anomalies. Their persistence over large areas and their coincidence with belts of shallow earthquakes are most striking facts. On the other hand, a hypothesis that starts with the forces and explains the observations would be much more trustworthy. However, no unimpeachable force has been found, and it seems very likely that several larger sources of stress differences as well as a large number of smaller ones cooperate in creating the changes that we observe.

E. Kraus[32] has tried to apply the hypothesis of subcrustal currents to the Alps. He considered the material to have moved downward along two vertical zones. His mechanism thus corresponds to the gravity observations that we have from regions with present tectonic activity and explains the conclusions of seismologists that the roots of the Alps extend downward much farther than the visible Alps project above sea level.

Hess[33] has pointed to the fact that in many regions, *e.g.*, in the West Indies, the East Indies and the Alps, serpentinite intrusions are correlated in a similar way with the negative anomalies. He supposes that in the Alps and in a part of the West Indies the down buckle causing the negative strip has been dissipated by fusion, isostatic uplift or both, so that now anomalies greater than -50 milligals are rare and the field of negative anomalies has broadened considerably.

According to Haarmann,[34] cosmic events produce subcrustal currents. The corresponding vertical movements at the surface are the *primary tectogenesis*. They are followed by *secondary tectogenesis* with a tendency to restore the equilibrium. Although the primary causes in Haarmann's theory have been considered improbable, the main idea has been developed by van Bemmelen, Grabau[103] and others.

Van Bemmelen[35] assumes that the original material of the earth was not absolutely homogeneous and that the differentiation in the primeval melt was influenced by local conditions as the main cause of tectonic activity. Later, further differentiations produced differences in density. Subcrustal currents have a tendency to restore isostatic equilibrium at depth; however, they produce geosynclines at the surface. Changes in pressure and differences in cooling add to the existing differences.

Under the central and deepest part of the geosyncline the activated magmatic differentiation will cause accumulations of salic and simatic products. When these accumulations have reached certain dimensions, the hydrostatic equilibrium will be disturbed to such a degree that an uplift of a median ridge will result. This is the embryonic stage of an orogenic cycle. An example

is the mid-Atlantic ridge. This median ridge will be volumetrically compensated by zones of subsidence on both sides. It may be the embryo of a cycle of laterally shifting meso-undations, because magmatic differentiation is also stimulated under the adjacent trough deeps. The impulse to undatory uplift will shift from the central axis of the geosyncline to both sides until the further development of the cycle is obstructed or checked by the continental forelands.

Van Bemmelen has discussed the details of his theory for a number of examples.[36]

An attempt to treat the undation theory and similar problems by mathematical means has been made by van Bemmelen and Berlage.[37] The plastic deformations connected with such problems have been studied by Bijlaard.[38] S. W. Tromp,[39] too, considers undulations as the major feature in mountain building. However, he disagrees with the theories mentioned thus far and believes that Helmholtz waves in the discontinuity between sial and sima control the movements in the earth's crust. He assumes that they are produced and maintained by the forces which shift the continents and by movements of the poles. According to Tromp, in the Cordillera of North America the amplitudes of the waves are 5 to 10 km., their wave length 1,200 km.; in the East Indies the amplitudes are 5 to 6 km., the wave length about 600 km.; the velocity of the waves probably exceeds 30 m. per year. Tromp refers to experiments that he has made. Serious doubts have been expressed about the data and the theory. No appropriate physical theory has been worked out for Helmholtz waves in a viscous solid.

E. Reyer,[40] M. Reade[41] and C. G. S. Sandberg[42] have assumed that thermal processes and sedimentation combine in producing mountains. However, the effects considered by them are too small.

Subcrustal currents due to the differences in temperature and properties of the crustal material have been studied by C. L. Pekeris.[43] Under certain assumptions he finds convection currents in the mantle beneath the crust, which, in the upper part of the mantle, are directed from the region under the continents toward the bottom of the oceans; in the central part of the oceanic region they turn downward toward the core, turn back toward the region with continental surface in the lower part of the mantle and ascend under the central parts of the continents. The speed depends on the conditions. Under various assumptions, Pekeris found maximum velocities between a fraction of a centimeter per year and a fraction of a meter per year. The positive isostatic anomalies over the oceans observed by Vening Meinesz and others, of the order of 30 milligals, probably could be brought about by

a proper assumption on the temperature perturbation (see also Ref. 96).

In addition, subcrustal currents due to thermal causes have been considered by Bucher.[7] "Fluctuations in the heat content of the subcrustal body of the earth constitute one of the factors which control the alternating contraction and expansion of subcrustal matter. Since Archeozoic time the heat content of the crust has decreased materially." The alternating swelling and shrinking of subcrustal matter cause alternately tensile and compressive stresses in the crust (*op. cit.*, page 477). Bucher discusses the mechanism of both contraction and expansion.

> The presence of . . . belts of excessive negative anomalies, the dominance of vertical uplift in the orogeny of the immediate geological past, general emergence of all continents and regression of epeiric seas, all show that the present represents a late stage in a compressional phase of crustal deformation. It is but natural, therefore, that we should be unduly impressed with the part which compression plays in diastrophism. (*Op. cit.*, page 482.)

As a whole, the importance of subcrustal currents for tectonic processes is pointed out in most of the hypotheses assuming movements of larger parts of the earth's crust; these hypotheses will be discussed in a later section of this chapter.

As a summary of the hypotheses that involve subcrustal currents we may state that the existence of such currents is highly probable, that they may be produced and maintained by thermal, chemical and mechanical processes, but that the relative importance of these processes is thus far not known either in general or in special instances.

CONTRACTION THEORIES

At least since the time of Descartes, the theory that the earth is cooling, shrinking and thus forming a crumpled crust has been considered. Although the hypothesis has been advanced repeatedly, *e.g.*, by Joly (see Chap. VII), Halm[53] and Lindemann,[44] that the earth is really getting warmer and expanding at least most of the time, it is generally held that the earth is cooling. There is, however, no agreement at all about the amount of cooling or as to whether this cooling is sufficient to produce a contraction large enough to explain generally the formation of mountains.

A method to calculate the contraction of the earth regardless of the cause of the contraction has been suggested by Meyermann.[101] His calculations are based on the times of ancient and modern eclipses. He finds that the change of the length of a day is smaller than would

follow from the tidal friction, and concludes that the radius of the earth is decreasing by about 5 cm. per century.

R. A. Sonder[45] gives a detailed discussion of the contraction theory. Three conditions must be fulfilled to explain mountain formation by the contraction theory: (1) The earth is contracting. (2) The crust is able to slide over the substratum. (3) The layers affected by the tectonic processes are able to accumulate the stresses needed to produce mountains. Three criteria indicate whether the theory is correct or not: (1) The amount of material now stored in the mountains must be of such an order as to agree with the possible contraction. (2) The order of magnitude of the stresses must be correct. (3) The sequence of orogenies in time and space must be appropriate to the theory—they must not be too close. Although the adherents of this theory agree that these criteria are met, a large number of geophysicists and geologists disagree and believe that contraction can explain at best only a fraction of mountain formation.

The physical side of the contraction theory has been investigated, for instance, by H. Jeffreys.[5] For this theory laws 6 and 7 mentioned at the beginning of this chapter are of fundamental importance. Law 6 corresponds to the second condition of Sonder, just mentioned, and if the suppositions are correct, proves it. Thus far there has been no objection, and the mechanism of the contraction theory is considered correct by most geophysicists: Stresses due to the contraction of the earth accumulate in the earth's crust until, finally, somewhere in a weak zone (geosynclines) the stresses reach the strength of the rocks, "complete fracture takes place in surface rocks, and the strength is reduced to zero. Thus crumpling continues until the stresses are almost completely relieved. This corresponds to an epoch of mountain formation. Then the fractures become sealed up afresh, and further interior cooling recommences the process" (*op. cit.*, page 138).

Jeffreys believes that the contraction theory accounts for at least the greater part and possibly the whole of mountain formation. A. Holmes's[46] objection he considers as the most serious, *i.e.*, that as the cooling of the earth decreases, this should produce an increase in the intervals between the orogenic periods, whereas the contrary seems to be correct. However, this may be due to the better possibilities for investigating their occurrence in the more recent periods. According to Jeffreys,[47] "at least the theory is not obviously wrong and remains the best available until some alternative can be shown to fit the facts as well or better."

Many others agree that the contraction theory has considerable advantages but still feel that there are certain doubts which have not

been removed. Stille[48] has discussed it favorably. Daly[49] has based his "down-sliding hypothesis" partly on the contraction theory, but he has assumed, in addition, the effects of changes in the speed of rotation of the earth and of erosion. The "orogen theory" of Kober[50] and the ideas of Ruud[97] are special forms of the contraction theory.

Though in the publications just mentioned contraction was considered the dominating process in orogeny, others have arrived at results indicating that the cooling of the earth is not sufficient to produce the major part of the crumpling. For example, under the reasonable assumption that the contraction of the radius of the earth does not exceed 1 mm. per century, Gutenberg has calculated that in the last 100 million years, or the post-Triassic time, only a mountain range 2 km. in height and 200 km. wide could have originated in this way. Under assumptions discussed by Geszti,[3] the mass of the Alps would correspond to a shortening of the radius of the earth of about 2 km. Thus, cooling during about 200 millions of years would have been needed to produce the contraction necessary to form the Alps.

There are two circumstances which have reduced the probability that the contraction theory proper is sufficient to explain all mountain making. One is the discussion of radioactivity which has resulted in the belief that the cooling of the earth is less than it had been originally believed. The other is that the cooling of the earth is decreased by the contraction. The potential of a homogeneous sphere referred to its center is $-0.6kM^2/r$, where M = mass, r = radius and k = constant of gravitation. A decrease in the radius of a homogeneous sphere of the size of the earth by 1 mm. produces a decrease in the potential energy corresponding to 8.3×10^{21} cal. This heat retards the cooling. Calculations of its amount in the earth have been made by O. Schmiedel,[52] who found, under the supposition that contraction is restricted to the outermost tenth of the earth, that about 40 per cent of the heat of the earth lost from radiation is restored by the corresponding contraction. This value is probably too high since the effect has probably been important in less than one-tenth of the planet. However, the correction to the amount of shrinkage may be taken as at least 10 and possibly even 20 per cent.

Considering these two causes for retardation in the cooling of the earth, Gutenberg came to the conclusion that the thermal contraction of the earth explains the formation of mountains in part but that other forces of the same order must have cooperated.[51] A similar point of view is held by v. Bubnoff,[89,90] Salomon-Calvi,[55] Schwinner, Kossmat, Watts[56] and others.

E. M. Anderson's[57] results point in a similar direction. After a detailed discussion he found, for example, that the loss of heat needed

to explain the formation of the Tertiary mountain ranges is about five times the loss which is calculated for the time interval between the late Paleozoic mountain building and the formation of these mountains. "The explanation of Tertiary mountain ranges afforded by the thermal form of the Contraction Theory is thus seen to be unsatisfactory. This result is more significant as the data chosen were extremely favorable to the theory" (*op. cit.*, page 158). In a second part of his paper, Anderson discussed the problems of elasticity and viscosity involved in the contraction theory and concluded that the results inferred in this way appear to tell against the validity of the theory. However, this conclusion should be combined with his statement that "the resistance of rock-masses to very long-continued pressures is at present unknown."

Objections of Holmes[46] have already been noted. He discussed in detail his statement "that the hypothesis of a continuously cooling earth had consistently failed to lead to any adequate explanations of fissure eruptions and other volcanic and tension phenomena, mountain building processes and their distribution in time and space, and the alternation of marine transgressions and recessions." Holmes suggested, instead, a "hypothesis of magmatic cycles." "The excess heat generated within the earth by the radioactive elements is accumulated for a time in the formation of magmas, and afterwards discharged by the ascent of the magmas to higher levels and their ultimate extinction by relatively rapid cooling" (*op. cit.*, page 276).

F. Nölke [Refs. 1 (pages 75–117), 58] stresses the fact that the contraction theory well explains the observed facts; on the other hand, he holds that the amount of contraction is not determined merely by the flow of heat through the crust. He thinks that contraction is due to intratelluric processes, which are either of thermodynamic character and only slightly or not at all connected with loss of heat, or possibly to changes in state or in atomic constitution, to degassing of deep-seated magma or to eruptions which cause decrease in volume of the deep interior of the earth. According to Nölke, the core of the earth is a specially important locus of further contraction.

In summary, we may state that the thermal contraction of the earth probably explains mountain building in part, but at present the evidence is rather in favor of the assumption that other processes play at least equally important and probably more important roles.

CHANGES IN CLIMATE AND MOVEMENTS OF THE POLES

The findings of paleontology and paleogeography[59] leave no doubt that considerable changes have taken place on the surface of the earth during its history. These sciences provide data regarding the distribution of land and sea as well as the climate. If we can find the

sources of these changes, important information on the interior of the earth may also be found. This fact as well as the unusually large degree of interest in these problems is the reason why an attempt has been made here to present in greater detail the hypotheses that have been offered to explain the changes in climate and in the distribution of land and sea.

So far as climate is concerned, the first reliable data concern the Algonkian. Evidence of glaciation has been found in the Great Lakes region, northern England, India, Australia and South Africa; however, in most cases the exact age cannot be determined. From the available evidence, it appears that during the Algonkian long periods of mild climate were interrupted by periods of glaciation. The Cambrian strata give indications of glaciation as well as of deserts, but the data, here also, are too scanty to give even a general idea of the climatic zones. The same is true for the Ordovician, the Silurian and the Devonian, except that there are good indications that the climate was then warmer in Europe and parts of North America than it is today and that during the Devonian the position of the climatic zones in the continental areas differed considerably from their present position.

Relatively good evidence is available for the Carboniferous and the Permian. Köppen and Wegener have tried[60] to reconstruct the climatic zones. There are indications of arid regions in North America, Europe and northern Africa and clear proof of extensive Carboniferous and more localized Permian glaciation in parts of South America, South Africa, Australia and India. It is difficult to find a position of the South Pole that would bring all the glaciated areas into the polar region, if we assume that the distances between the continents were as they are today. This is one of the arguments in favor of the hypothesis of continental displacements, a hypothesis that would seem to solve the problem, provided that one also assumes wandering of the poles. Accordingly, Köppen and Wegener and others place the North Pole of the Carboniferous in the northern Pacific Ocean northwest of Hawaii and the Permian North Pole somewhat more to the northeast. The corresponding positions for the South Pole are between South Africa and Australia. During these times, these continents were assumed by Wegener to have been close together. He had the Equator crossing Central America, southern Europe and Central Asia. Such a position of the poles had already been deduced (in 1902) by Kreichgauer.[61] Figure 8 is reproduced from his book. All who accept the general hypothesis of movements of the poles have found, independently, a path of the North Pole since the Carboniferous close to that given by Kreichgauer and agree with the placing of the

poles for the Carboniferous itself. However, there is no doubt that his assumptions for the preceeding periods are not supported by observations.

The data for the Mesozoic era are more scanty; there are scarcely any indications of glaciation.[62] The vegetation and the fossils seem to indicate relatively small changes, probably slow movements of the climatic zones toward their present positions.

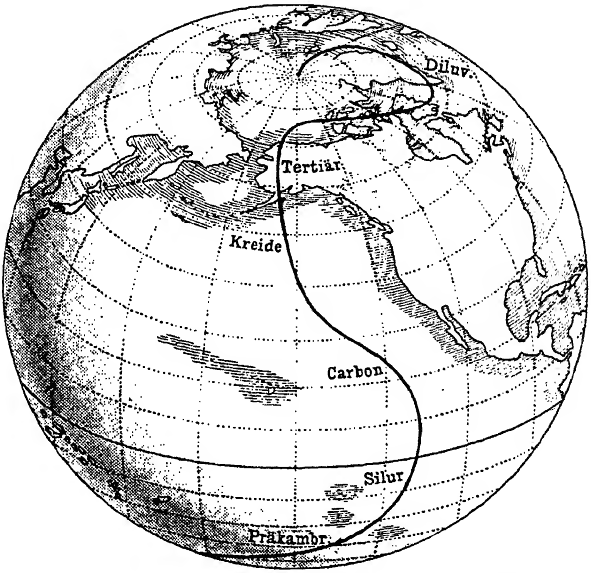


Fig. 8.—Path of the North Pole after Kreichgauer. The part preceding the Carboniferous is generally considered incorrect.

During the Tertiary, the climate changed more rapidly. That of northern Europe grew perceptibly cooler than it had been before and approached present conditions. In the Miocene, plants were still growing in central Europe which could not stand the cooler climate of the present time. On the other hand, indications point to a cooler climate than that of today for the Pacific coast of North America. During the Quaternary, present conditions were approached; detailed investigations favor a path of the poles as indicated in Fig. 8.

The characteristic events of the climate of the Pleistocene were the successive glaciations, separated by interglacial stages.[59,60,63,64,65,95] About four major and several minor glacial stages have occurred during the last 600,000 years; the last large minimum in temperature

was about 60,000 years ago, and the retreat of the ice from central Europe began about 40,000 years ago.

There is a marked difference between the relatively short glacial periods of the Pleistocene and the changes in climate during earlier periods. While the Pleistocene "ice age" consisted of a number of periods during which the temperature apparently was lower over large parts of both hemispheres, changes in climate during the earlier history of the earth are indicated by large areas with noticeably warmer climate than we have today, contrasting with other areas that at the same time had an unusually cool climate. For this reason it seems probable that the two phenomena have completely different causes. One produces more or less periodical changes of temperature which may have existed during the whole history of the earth; the Pleistocene glacial and interglacial periods are examples of such periodical changes. Other causes are responsible for the much slower changes in climate during periods as long as geological eras and which have simultaneously affected one part of the earth in one way and another in the opposite manner. Thus certain regions, such as Iceland or Antarctica, which are very cold now, for the late Paleozoic or the Mesozoic era show clear indications of what we would call subtropical climate today, but no trace of glaciation; at the same time other regions were at least temporarily glaciated. This fact can scarcely be explained by assuming world-wide cooling as supposed for the Pleistocene glaciations.

Explanations of climatic changes fall into two classes: (1) changes in the position of the poles are supposed, or (2) attempts are made to explain the climates without assuming such changes. A large number of suggestions of the second type have been discussed by C. E. P. Brooks.[67] He stresses the cooling effect of extended icecaps, arguing, for example, that an original fall in temperature from the neighborhood of the freezing point by 5°F. (3°C.) and subsequent formation of a large icecap in the Arctic region lowers the temperature by about 50°F. (27°C.) in high latitudes. If this fall in temperature is large enough, a high-pressure area will develop, a "glacial anticyclone," with clear skies and intense cooling by radiation. "The outwardly-directed winds spread Arctic conditions in a broad zone round the margin of the ice, and may even result in the 'sympathetic' glaciation of a neighbouring mountain range. It is probable that the great development of Alpine glaciers during the Quaternary was due partly to cooling by the winds blowing off the Scandinavian ice-sheet" (*op. cit.*, page 70). The fact that the ice reflects more heat than do rock or water increases the cooling effects.

Ocean currents are another important factor. The role the Gulf Stream has played in the climate of western Europe and Iceland is well known. In his book Brooks has discussed papers published on the question of changes in the Gulf Stream during the Quaternary. Though the changes in Central America probably affected it little, "the closing of the gap between Greenland and Europe by the elevation of the submarine ridge which passes through Iceland . . . to Scotland, which occurred during the Quaternary Ice-Age, must have deflected the Gulf Stream" and probably increased the severity of the glaciation in the countries bordering on the north Atlantic (*op. cit.*, page 90).

Other factors of importance making for changes in the climate, for which Brooks has discussed literature available in 1925, are changes in the radiation of the sun[68] and changes in the absorption of radiation by the atmosphere, due, for example, to varying content of carbon dioxide, volcanic dust or water vapor (clouds). Such changes would affect the earth as a whole.

The effect of the "continentality" on the climate has been discussed in detail by Brooks (*op. cit.*, pages 147–179). He found the following changes in the temperature produced by the presence of 10 per cent land (a) to the west, (b) to the east and (c) of 10 per cent ice for a point in a given latitude φ, in degrees centigrade:

	January			July		
φ	a	b	c	a	b	c
70°N.	−4.3	−2.0	−4.6	0.2	0.2	−1.6
60°N.	−3.1	−0.1	−0.7	−0.1	1.1	
50°N.	−2.9	0.9	−0.9	0.4	0.6	
30°N.	−0.8	0.3	.....	0.8	−0.1	
0°N.	0.1	0.0	.....	0.2	−0.1	
20°S.	0.7	0.0	.....	0.2	−0.2	
40°S.	0.9	−0.3	.....	0.0	−0.3	

These data are based on the present distribution of land and sea. The effect of land combines with the effect of elevation. This is due to two major causes: the lower temperature because of the higher altitude and the effect on the precipitation, which in itself is another cause for changes in climate.

Finally, Brooks discussed the climates of the past and considering the various factors gave the following formula (*op. cit.*, page 245) "for calculating the mean temperature between 40° and 90° north latitude in any part of the Mesozoic or Tertiary;

Temperature (°F.) = 48 − 0.13 × Continentality (%) − 0.45 × elevation (hundreds of feet) + 0.43 × Ocean Currents (%) − 0.26 × volcanicity (arbitrary scale 0–10)."

For earlier periods the calculated and the observed values differ more and more, the farther back one goes.

There is one more group of factors affecting the climate that has been mentioned by Brooks: changes in the obliquity of the ecliptic, changes in the eccentricity of the earth's orbit and the precession of the equinoxes. The effect of these factors has been studied in detail by Milankovitch [Refs. 60, 69, 70 (pages 207–214)]. He found, as a first approximation, the following equation for the total radiation Q in one half year (either summer or winter):

$$Q = W \Delta\epsilon \mp m \Delta(e \sin P). \tag{18}$$

W is a constant that depends on the latitude of the point under investigation and has one value for the summer and another for the winter. $\Delta\epsilon$ is the change in the obliquity of the ecliptic against the present value of 23° 27′ 30″; m is a constant depending only on the latitude; e is the eccentricity of the earth's orbit; P is the longitude of the perihelion. $\Delta(e \sin P)$ is the difference between $e \sin P$ for the year under consideration and its present value. The sign minus between the two terms of the equation is to be used for the northern hemisphere in summer and the southern hemisphere in winter; the sign plus in the two other cases.

An increase in the obliquity of the ecliptic decreases the differences between the latitudes and increases the differences between the seasons. It has the same effect on both hemispheres. The changes seem to have a period of 40,000 years. P has a period of about 21,000 years; during this time $\sin P$ varies between +1 and −1. e has a period of 92,000 years; its amplitudes are between 0 and 0.0677. If $P = 90°$, the differences between the summer and winter are a minimum on the northern hemisphere and a maximum on the southern; for $P = 270°$, the case is just reversed.

The values of the constants involved are well known. Supposing that there has been no change, it is possible, according to Milankovitch, to extrapolate backward for about 600,000 years without a probable accumulation of the errors of more than 10 per cent of the maximum amplitudes of Q. In his earlier papers[60,69] Milankovitch used data given by Pilgrim (1904) and based on results of Stockwell (1873); for his more recent calculations[70] he has used revised values given by Michkovitch based on the results of Leverrier (1873). The

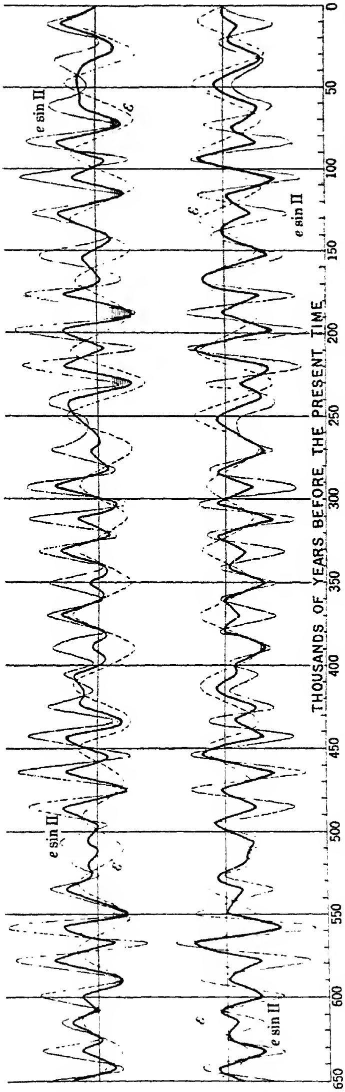


FIG. 9.—Secular changes in the heat received at 65° latitude (upper half, northern hemisphere; lower half, southern hemisphere) during the summer (heavy curves). The curves are calculated from the changes of the obliquity ε of the ecliptic, the eccentricity e of the earth's orbit and the longitude Π of the perihelion. (*After M. Milankovitch.*)

differences do not affect the general results, although they are large enough to be seen when the curves are compared.

In Fig. 9 the light line gives $e \sin P$ from the earlier paper[60] of Milankovitch, the broken line the value of ϵ. The heavy line of the upper curve is half the difference of these quantities, corresponding approximately to the radiation during the summer of the northern hemisphere, the heavy curve of the lower figure half their sum, corresponding to the total radiation received during the summer of the southern hemisphere. For more accurate data, the constants W and m must be used for a given latitude. This has been done in the curves reproduced in the upper part of Fig. 10. Milankovitch assumed—following the example of other climatologists—that heat received during the summer is much more responsible for the climate of a cold

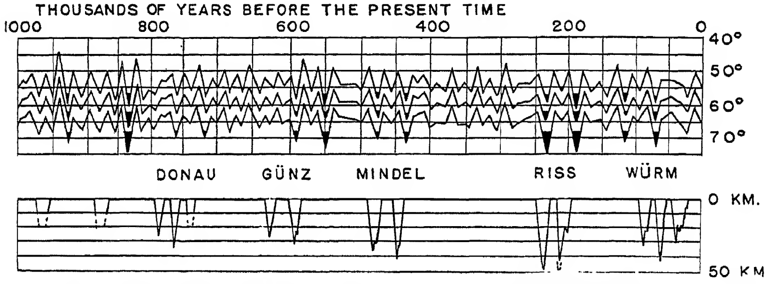


FIG. 10.—Upper half, changes in climate as calculated by M. Milankovitch. The three curves correspond to a point at 55°, 60° and 65° north latitude respectively. Lower half, distance of the end moraine of the Alpine glaciation from the Alps. (*After Eberl.*)[66]

region than the heat received during the winter. Great cold in winter does not affect the possibility of glaciation so much—the ground is frozen in any case—as does coolness during the summer. To make the results clearer, the curves given in Fig. 10 represent, not temperatures, but the latitude that now receives the equivalent heat of the northern hemisphere during summer. For example, the uppermost of the three curves corresponds to a point at lat. 55°N. If we follow the curve from the right end (today) backward, we find that the calculated heat received by this point during the summer half year (183 days with maximum radiation) about 10,000 years ago was about the same as that received today by a point in 50°N. during the summer, and thus the climate was relatively warm. On the other hand, the summer heat about 70,000 years ago was much less and corresponds almost to the summer heat of a point now 5° more to the north. Milankovitch correlates this period with the last large stage of the Würm glaciation.

In a similar way, the central curve represents the heat received during the summer by a point in 60°N., and the lower line, one in 65°N. The lower part in Fig. 10 shows the distance of the end moraine of the Alpine glaciation from the Alps as given by Eberl.[66] The agreement is very striking.

If the explanation of the recent glacial periods as worked out by Milankovitch is correct, the southern hemisphere must have had glacial periods about the same time as the northern hemisphere, but the amplitudes should differ, and the peaks of glacial stages on the two hemispheres should not coincide.[93]

The main objection raised against the hypothesis is that glacial periods of the type here discussed should have occurred during the whole history of the earth whereas the climatologic evidence in Europe does not show them. There are various possibilities that explain this fact. One is that the values involved in the calculation were different in the earlier periods. The other, which seems much more probable, is that they always have existed, perhaps with larger, perhaps with smaller amplitudes than now, but that other causes had produced a warmer climate in Europe and in other regions which did not permit glaciation, even though the radiation during the summer was reduced by an amount equivalent to a shift in latitude by 5°. Search for such periodic changes in climate either must be guided by evidences other than signs of glaciation in rocks or must be turned to regions where old glaciation has been proved.

The most promising periods for such investigations are the Carboniferous and the Permian. A thorough study of the literature on glaciation during these periods has been made by Salomon-Calvi.[71] According to the findings of David and Süssmilch,[92] in Australia at least five and possibly six major glacial periods have existed, separated by long nonglacial periods. For other regions at least two have been identified, but certain observations indicate a larger number. The present writer agrees with Salomon-Calvi that these results indicate periodic changes in radiation which correspond to the theory of Milankovitch and which are superimposed on a change in climate of long duration. Probably, such changes in climate, according to the theory of Milankovitch, have existed during most of the earth's history, but only during the Pleistocene, the Carboniferous and the Permian did they produce periods of glaciation in regions where they can now be investigated from geological or paleontological evidence.

The problem now remains how to explain the changes of climate on a large scale as between the climate of the Carboniferous period and that of today. It may be possible to apply a combination of causes

discussed by Brooks. It is also possible to assume that animals and plants reacted in a different way to climatologic conditions than they do at present, so that, for example, palm trees could grow in central Europe at a time when the temperatures on the earth were so much lower that large parts of South Africa were glaciated, although this seems unlikely to the present writer. A different composition of the atmosphere may have produced great changes. However, a good many authors quite independently have found that there is no difficulty in plotting climatological zones for a given period, which leave little doubt about the most probable position of the poles. The fact that their findings are all nearly the same for the periods since the Carboniferous indicates that the drawing of these zones is not arbitrary. The major question is how good the material is in itself. However, many revisions in interpretation are needed to destroy the evidence of the changes in position of the climatologic zones in the course of time. If we so believe, we come to the hypothesis that the poles have moved.

This does not mean that the axis of the earth has shifted inside the earth—an extremely large amount of energy would be needed to produce such a movement. Nevertheless, no very large forces are needed to shift the surface layers of the earth over the plastic interior. Possibly, the *Polfluchtkraft* would be sufficient to produce this effect, and Milankovitch has found[72] that the path of the poles theoretically resulting from this hypothesis follows closely the path deduced from the climatologic observation. The crust of the earth would have moved, in this case, from an unstable position in the Carboniferous with the north pole at about 168°W. and 20°N., first slowly toward the region near 150°W. and 40°N., then faster, following about 145°W. into its present position. The point toward which it should move in the future with a decreasing speed is at about 50°E. 65°N. Of course, this theory presupposes that the distribution of mountains, land and ocean always has been the same and will not change in the future. The question how the earth's crust came into the unstable position that it had during the Carboniferous, according to the hypothesis, is easily answered by the fact that doubtless at that time the distribution of continents and oceans was quite different and that possibly the crust had reached this position in a movement toward equilibrium under the surface conditions existing at a time prior to the Carboniferous. Naturally, the formation of mountain chains requires much larger energies from another source.

A more serious question at present is why we do not observe movements of the pole corresponding to those which have been found as an

average during the recent past and which amounted to the order of 1° in 10,000 years. The motion of the poles as calculated from observations during the most recent decades would correspond—if continuous and if not influenced by local displacements of the observatories—to about 1 min. during 10,000 years in the direction toward Greenland.[73] The movement found from the changes in climate for the North Pole during recent time would have had the opposite direction. However, periods of slow and reversed movements have been inferred from the observations of climate for other periods. Nevertheless, there is a lack of confirmation. Another difficulty has been added by the conclusions in regard to deep-focus earthquakes (Chap. XI). The foci of the "intermediate shocks" which occur at depths of the order of 100 to 200 km. are below the Tertiary mountain belts. If we were certain that they are still continuing along the old faults of the Tertiary, such knowledge would disprove the theory of the movement of the continents relative to the substratum, which has been inferred from the climatic changes. However, it is not improbable that these shocks occur in a zone of plastic flow (Chap. XI). In this case, the old faults were completely destroyed, as the flow very probably has quite different velocities in different regions of the same fault. The present intermediate shocks must be produced by stresses that are connected with mountains and must create new faults within time intervals that are of a smaller order than the geological periods. A similar interpretation would hold for true deep-focus earthquakes which occur around the Pacific Ocean.

A great number of theories have been advanced in which movements of the poles, or, better, movements of the crust relative to the axis of the earth, have been assumed. Kreichgauer's[61] conclusions have already been mentioned (see Fig. 8). Köppen and Wegener[60] have found a similar path for the poles since the Carboniferous; for the Pleistocene ice age they have adopted the explanation of Milankovitch. They have, moreover, supposed that Wegener's theory of the continental drift is correct; this theory will be discussed in the next section of this chapter. Gutenberg[74] agreed in general with the results of Köppen and Wegener, except that he replaced Wegener's idea of the drifting apart of the continents with his hypothesis that they flow apart. Figure 11 shows the position of the continents at various periods with the major climatologic evidence on which the assumptions are based. Sketches for the Carboniferous-Permian have been made by Salomon-Calvi[71] and by Du Toit.[88]

Movements of the earth's crust relative to the axis of the earth must be accompanied by vertical displacements. A block with a

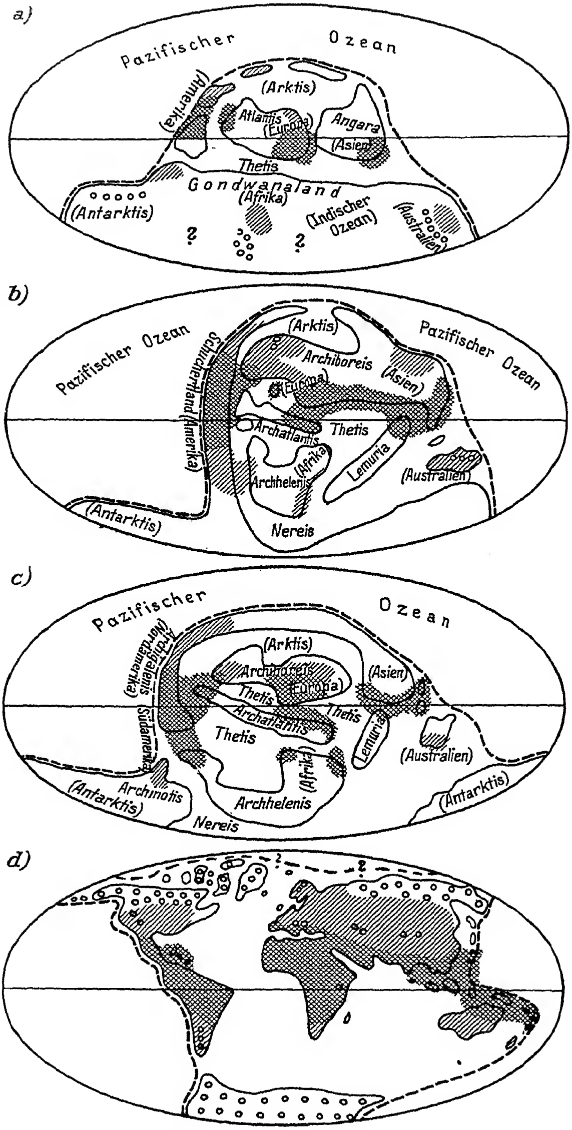


FIG. 11.—Probable distribution of continents and of climates; after Gutenberg (1927). *a*) Carboniferous; *b*) Cretaceous; *c*) Eocene; *d*) Recent. Circles indicate cold climate; single shading, moderate climate; cross hatching, tropical climate. Probable boundary between region with a sialic crust and that without sial is marked by a broken line.

thickness of 50 km. in equilibrium near the equator should have a thickness of about 49.6 km. near the poles to be bounded by the same equipotential surfaces there. If it moves toward a pole, it must sink deeper to keep in equilibrium. As this does not occur at once, on a block moving poleward we must expect too high gravity. As the ocean represents the true level of equilibrium and the block drifting poleward is too high, regressions must be observed there. In other words, regressions are to be expected in regions toward which poles seem to move, transgressions in regions from which the poles are moving away. Wegener investigated this problem and found that, in general, the observed transgressions and regressions during several geological periods agree with those to be expected from the assumed movements of the crust or poles. However, further investigations are needed to decide whether or not these conclusions are correct. The problem is much complicated by the fact that due to the variations in the icecaps of the poles the level of all oceans may change considerably all over the world in the same direction (see Chap. III).

Other hypotheses supposing movements of the poles include, for example, that of Simroth,[75] who has assumed that the poles oscillate with decreasing amplitudes, and of A. Heim,[76] who has considered movements of the poles to be due to cosmic impulses combined with accelerations in the velocity of the rotation of the earth. A decrease in the speed of rotation of the earth due to tidal friction (partly combined with polar movements) has been made the basis of hypotheses by Böhm,[77] Blytt[78] and Quiring.[79]

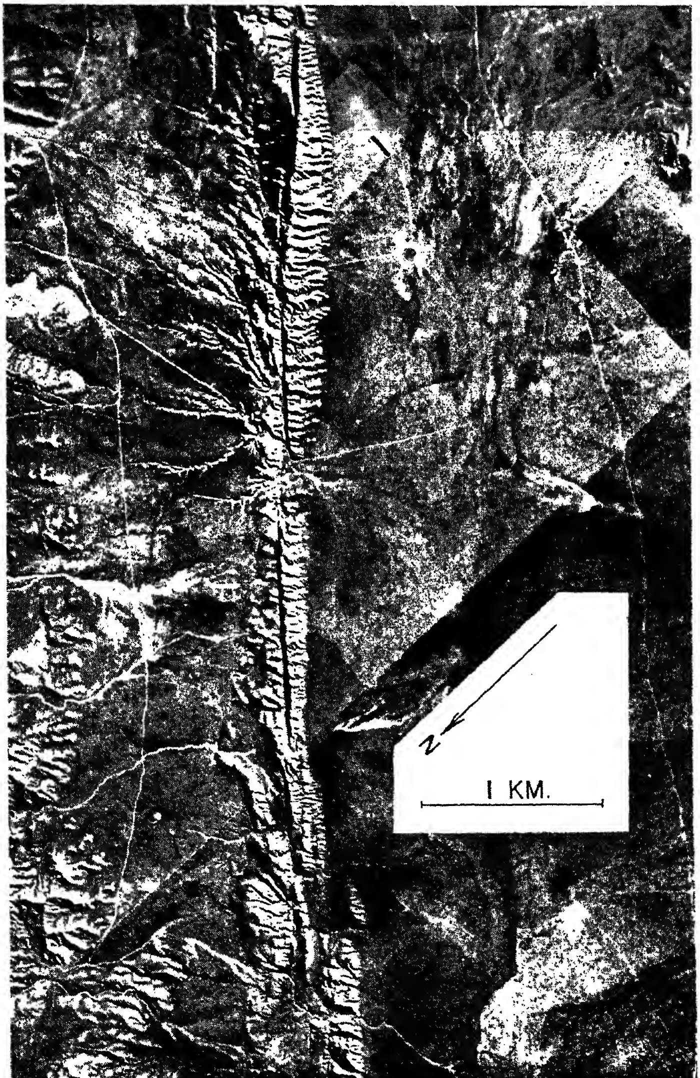
In reviewing the hypotheses of the last section, there can be no doubt that the observed changes in the astronomical elements of the earth's orbit affect the climate noticeably and that long periods with cooler or warmer summers must result in this way. Whether or not these changes are sufficient to produce glacial periods has not yet been decided. The surprisingly good agreement between the theory of Milankovitch and the observations for parts of the northern hemisphere make it very probable that the Pleistocene ice age was due to such changes in astronomical elements. Glaciation was increased by the fact that the outflow of cold air from a large area covered by ice decreases the temperature of the whole region. Possibly such periods of relatively cool summers have existed during most of the history of the earth, but observations from regions affected by glaciation are too scanty for other geological periods to permit of the investigation of glacial and interglacial stages, except for the Carboniferous and Permian, where similar periodical glaciations have been found. In regions with warmer climates, on the other hand, the differences are probably too small to be revealed by the study of plants.

These oscillations in climate of relatively short duration are superimposed on changes that take place during time intervals of the length of several geological periods. Whether they are produced by movements of the poles or by a combination of changes of the earth's crust (continentality, ocean currents, volcanic activity producing more dust, formation of high mountain ranges) with variations in the heat received at the surface of the earth from the sun or by a combination of all these factors is an open question.

MOVEMENTS OF LARGE PARTS OF THE EARTH'S CRUST RELATIVE TO EACH OTHER

If one assumes that changes in climate are to be explained by movements of the earth's crust relative to the poles, it is only a small step to the assumption that large parts of the continents move relatively to one another. The possibility of such movements is proved by the gradual changes in the earth's crust that can be observed. Horizontal shifting of large areas may be detected by astronomical observations (changes in latitude) or by comparison of time signals from far-distant points with the astronomically controlled time of the place under consideration (change in longitude); movements of blocks, one relative to the other, may be discovered by geological investigations; clear indications of vertical movements are often shown by changes in shore lines or by records of tide gauges; finally, accurate data on present changes are provided by geodetic measurements, either triangulations or precise levelings.[102]

Data from any of the methods mentioned are very scanty, and, even when the findings available from all are combined, there is no considerable region of the earth for which a map showing recent changes can be drawn. Moreover, in many instances the results that have been found have either been shown to be erroneous or at least have been suspected to be so. This is true for astronomical as well as for geodetic measurements. Several independent investigations on the distance between Greenland and Europe gave a consistent increase in distance of 9 m. per year as an average between 1823 and 1870 and of 32 m. per year between 1870 and 1907 for northeast Greenland; they also gave an increase of 20 m. per year between 1873 and 1922 and of 36 m. per year between 1922 and 1927 for the distance from Godthaab to Europe. In spite of the observation of a continuous shift of between 1 and 2 km. in a century, based on several independent measurements, the fact that no change has been found between 1927 and 1936 between West Greenland and Europe has been considered by Nörlund[80] as an indication that "most likely the deviations of the old observations from



... 12.—Offsets of streams along the San Andreas Fault at the Carrizo Plain, California (about 119½° West 35° North). (*Taken by the Fairchild Aerial Survey, Los Angeles, for the Barnsdall Oil Co., in February, 1936.*) Directions of the streams from left to right.

the new ones are the result of observation errors." Many geodesists have suspected this before, but others are convinced that at least the direction of the movements is beyond doubt. In no other instance has a change in distance been found between continents larger than the probable error.

However, considerable movements have accumulated during longer epochs. The horizontal displacements along the San Andreas Fault in California possibly amount to a noticeable fraction of a degree. Displacements of the order of many hundreds of meters during comparatively recent times are evident from offsets of streams (Fig. 12); accumulated displacements of the order of 40 km. have tentatively been suggested by L. F. Noble[81] from comparison of corresponding rocks on the two sides in its central part. In the northern section, the recent displacements seem to be much smaller.[98] The fact that along a large fraction of its extension no structures have been found that correspond to each other on the two sides indicates a relatively large displacement there. Since, apparently, the horizontal movements in a large number of other faults in central and southern California are of the same type—the block west of the fault has recently moved relatively northwestward[82]—and the tectonic picture of the Sierra Nevada possibly indicates a similar shear,[99] there can be no doubt that horizontal shearing movements extending over at least a large fraction of California have existed and still exist. Horizontal movements of the same order are known for other regions (see, *e.g.*, Ref. 56). Geodetic measurements in Japan have shown that at present horizontal movements exist there (gradual as well as in earthquakes) having a velocity up to 1 cm. per year,[83] which is of a higher order even than that characterizing the movements in California.

Large vertical movements are well known; the uplift of Scandinavia since the last glaciation (maximum probably between 250 and 300 m.) and in the region between the Great Lakes and the Hudson Bay (about of the same order)[64] are examples (see also Chap. XII). In all theories on mountain formation, relatively large horizontal and vertical movements must be assumed. It is only a short step to the conclusion that the distances between the continents have changed.

Most of the hypotheses dealing with changes in distance between continents or between parts of one continental block are based on observations. The assumption of "epeirophoreses," as Salomon-Calvi[55] has called such movements, in general explains certain facts better than any other hypothesis. The difficulties that are involved concern the forces as well as the mechanism.

No manifestly adequate force has yet been discovered. However, this is the weak point of all hypotheses that try to explain orogeny. On the other hand, it seems possible to develop a reasonable theory involving both subcrustal flow and relative displacements of crustal blocks. Since stress differences at greater depth must be accompanied by stress differences near the surface, and vice versa, it seems reasonable to assume that the forces which cause the large distortions of the crust are produced at a depth.

The second difficulty concerns the mechanism. The strength in the earth's crust is considerable (see Chap. XV) and does not permit plastic flow unless the stress difference rises to about 3,000 kg. per square centimeter. It is not easy to account for such large stress differences. Movements in the substratum, however, where the strength is probably smaller, do not require such large stress differences; but there, to be effective, they must long persist, for we know that the viscosity at depth is high (see Chap. XV).

The facts on which the hypotheses of large crustal displacements are based include movements that have been mentioned at the beginning of this section. In the preceding section we reviewed the climatic evidence for the displacement of the continents relative to the poles and to one another. There is an additional fact to be considered. Large parts of the continents, now well above sea level, were formerly submerged when other parts of the crust, it may be including those in their immediate neighborhood, remained above sea level. The reverse is also true: land bridges long existed across the Atlantic and Indian oceans. This is the conclusion of nearly all specialists in the field, although there is some disagreement about the extension of these bridges and their times of duration. The evidence is based on the spread and distribution of animals and plants. Land life was able to roam from land to land, although these lands are now separated by deep ocean water. The former continuity of continental areas prevented sea life from passing easily from one part of the ocean to another. That the new, separating areas of deep sea originated by simple downwarping or downfaulting of the crust is a hypothesis that contradicts the established principle of isostasy. The assumption of continental migration avoids this difficulty. But some paleontologists prefer to disbelieve in isostasy to the assumption of subcrustal currents, and some geodesists, for the same reason, assume that ocean currents carried spores, seeds or eggs and even live animals across the oceanic barriers in numbers sufficient to destroy the paleontological argument for continental migration.

A summary of criteria for continental drift has been given by Du Toit (Ref. 88, page 51). There are significant similarities in the structure of the continents on the two sides of the Atlantic and, in somewhat less striking degree, on the opposite sides of the Indian Ocean. Their correspondence cannot be said to prove continental displacements, but they must be regarded as highly suggestive.

The general idea of relatively large horizontal displacements of parts of the earth's crust is old. Long ago E. Suess[94] concluded that the mountain systems in central Asia are due to movements of the northern part of Asia toward the south and that the arcs of eastern Asia were produced by the push of that continent toward the Pacific Ocean. These ideas have been developed further by Taylor,[84] who believes that tidal forces were responsible for the movements.

The starting point of Wegener's theory[85] is quite different. Beginning with the similarity of the contours of the continents at opposite sides of the Atlantic and Indian oceans, he developed, since 1910, the hypothesis that the continents formed one block at the beginning of the Carboniferous and that they broke apart and drifted during the following geological periods, as indicated in Fig. 13. He discussed in detail the consequences of the breakup. Several hundred papers have since been published, partly in favor, partly opposed to the entire theory or phases of it (see, *e.g.*, Refs. 55, 56, 86, 87, 88). Some of the objections had their origin in Wegener's attempt to deal with subjects outside his own field of geophysics. Incompleteness of data and arguments was unavoidable in view of the fact that the synthesis involved problems of geophysics, geodesy, geology, paleontology, climatology, oceanography, volcanology and geochemistry. Wegener's was the first attempt to explain, in a comprehensive way, a wide variety of major facts. His argument was left with a weak link, one common to all tectonic theories; the forces he assumed are in part insufficient to produce the desired effect, while others, such as the hypothetical west-drift force, do not exist. Other objections include his assumption that the bottom of the Atlantic and Indian oceans are of the same material (sima) as the bottom of the Pacific, whereas geological as well as seismic data indicate that the upper layers are of the same type for the whole earth, except the Pacific and possibly parts of the Arctic basins (Chap. XII).

To remove these and other discrepancies in Wegener's theory without affecting its advantages, Gutenberg[74] has suggested a modification of the migration idea, by supposing that the continents did not break but flowed apart and that a connection of continental material still exists between them across the bottom of the Atlantic and Indian

oceans. Accordingly, the continents (see Fig. 11) were less widely separated at some earlier geologic times than they are at present, but

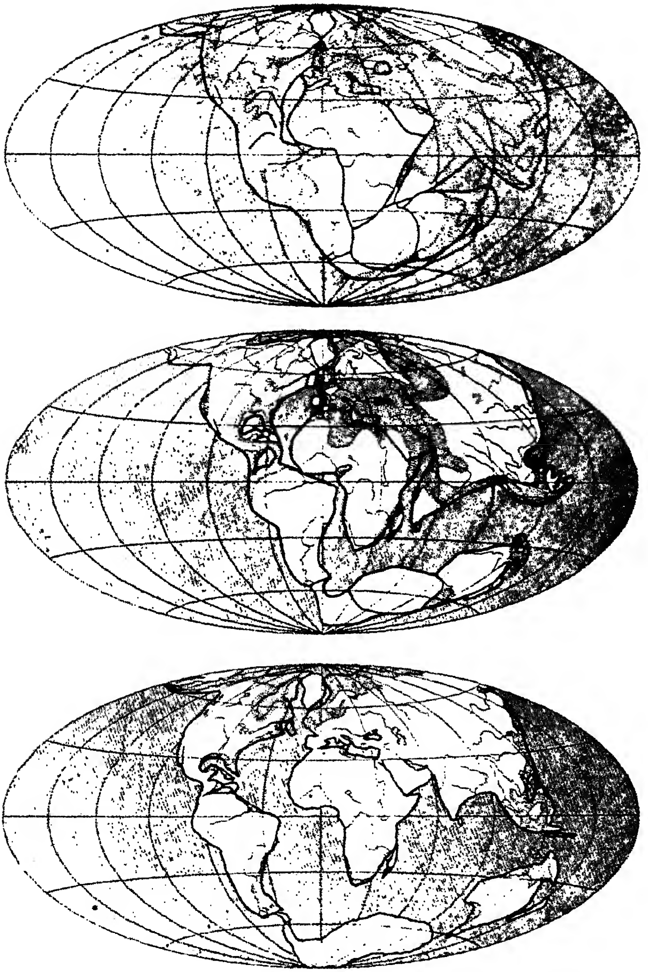


FIG. 13.—The drifting of the continents after Alfred Wegener. Top, upper Carboniferous; center, Eocene; bottom, older Quaternary. Dotted areas, shallow seas; present outlines, rivers and coordinate system only for purpose of identification.

the distances between the corresponding blocks were not less than about half the present gaps. At the time when two continents were

separated by this minimum distance, they formed one large continent without ocean between them. The possibility remains that the continents at some earlier time may have been farther apart than they are today. Whereas Wegener has assumed that the continents are drifting like icebergs, Gutenberg postulates plastic flow (including creep) in both the continents and the substratum (*Fliesstheorie*). The whole picture for each period is entirely based on observational data; the outline of the block with continental surface layer is derived mainly from seismic data (Chap. XII); it is supposed to be the same through the ages. The distances between the continents are based on the conclusions of the paleontologists as to the connections between the continental regions for the given period. The orientation of the poles and the Equator is chosen in such a way as to give the most likely explanation for the observed climatological evidences. The result is represented by Fig. 11. For the periods preceding the Carboniferous neither do the conclusions of paleogeography permit of finding the probable distances between the continental blocks, nor are the climatological evidences plentiful enough for the determinations of the location of the poles. However, the data fit much better in the sketch for the Carboniferous than in one for recent time; the possibility that practically the whole continental block was in the southern hemisphere is not excluded.

The sketches of Fig. 11 are only a first attempt of this kind. The figure should be revised by a specialist in paleogeography and paleontology. Discrepancies between observations and the sketches probably can be removed by a few small changes. Here too the most serious problem to be solved involves the forces that have produced, and may still be causing, continental displacements. The *Polfluchtkraft* seems to explain the general direction in which the continental blocks have been moving; however, this force seems to be too small, although the movement of each block as a whole over the substratum and the subcrustal currents in the opposite direction does not require large energies. But it does require much time. The spreading of the continents needs much higher stress differences to overcome the strength in the upper layers. Possibly the energy could be furnished by stresses resulting from the deviation of the earth's crust from hydrostatic equilibrium (difference in level between mountains and oceans). However, other forces are needed to produce these mountains. As has been pointed out repeatedly, such forces must exist, but we do not know them. They may be one of the types discussed in connection with subcrustal currents [*e.g.* Refs. 71 (page 143), 87, 89].

S. v. Bubnoff[89] and Du Toit,[88] following the ideas of Joly and Holmes, have pointed to the possibility that a part of the heat generated in radioactive processes may produce convection currents and thus furnish the energy for movements of large parts of the earth's crust. A hypothesis of continental movements based on geologic evidence (as any such hypothesis must be[90]) and supposing thermal energy is considered by these writers as more likely than any other hypothesis, as it would explain orogenesis as well as the related problem of petrogenesis. Similar ideas have been expressed by Watts[56] and Kirsch.[100]

Another hypothesis relying on continental movements has been suggested by R. Staub.[91] He assumes that the bottom of the Pacific Ocean is stronger than the continental material which consists of the *Laurasia block* (northern continents) and the *Gondwana block* (southern continents), which were separated by the Tethys. Under the action of the various forces that change their direction because of changes in the location of the acting masses, the two continental blocks approach each other, forming mountains along their battle front, the Tethys, and separate again. The theory would explain the European-Asiatic mountain chains, especially the Alpide system, but neither its failure to solve the problem of the required energy nor its neglect of the American mountain systems looks very promising for a development of this hypothesis.

To summarize the theories concerning large movements of continents relative to each other, we find that they intend to explain the changes of large parts of the earth's crust which must be assumed to fit maps found in paleogeography; they must be in agreement with the conclusions of geophysics in regard to the structure of the elements of the earth's crust. Their main weakness is the fact that thus far they fail to give us a full understanding of the forces that could produce these changes.

SUMMARY

The earth's crust is and always has been under the play of many forces. Chemical changes, differences in temperature in various parts of the crust and differences in structure, cooling of the earth, heat produced in radioactive processes are probably the major and the primary factors that have jointly controlled its development and have produced contraction, subcrustal currents and shifting of parts of the surface relative to each other and to the poles.

The following outline gives some of the processes that the present writer considers probable:

Cause	Result
1. Chemical processes; magmatic differentiation (combined with gravity) Radioactivity	1. Formation of the core and the layers of the mantle and of the crust Volcanism Subcrustal currents Mountain ranges
2. Cooling of the earth; other intratelluric processes	2. Contraction, formation of mountain ranges
3. Differences in temperature and thermal properties as between continents and layers below the oceans	3. Small subcrustal currents
4. Sedimentation	4. Shelves; geosynclines; small subcrustal currents
5. Erosion	5. Rising of region; small subcrustal currents
6. Changes in the astronomical elements of the earth's orbit with periods of a fraction of the geologic periods	6. Periodic changes in climate; glacial and interglacial stages
7. Movements of the continents relative to the axis of the earth [due to subcrustal currents (see Results) or differences in structure of the crust]	7. Changes in climate. Regressions and transgressions
8. Difference in structure between the continental block and the Pacific basin; extended and high mountain chains	8. Spreading of the continents; subcrustal currents; *Polflucht* of the continents(?)
9. Subcrustal currents (see Results, especially item 1.)	9. Changes in the crust; displacements; formation of mountain ranges
10. Large changes in ice load	10. Vertical movements; small subcrustal currents. Changes in sea level.

The relative importance of the processes just mentioned is still very doubtful. Subcrustal currents controlled mainly by thermal processes (radioactivity) seem to play the most important role. More definite results will lead to better information on the interior of the earth. For this reason, some space has been devoted to the hypotheses on the development of the earth's crust, which concern, without doubt, some of the most important outstanding problems of the physics of the earth.

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CHAPTER X

EVIDENCE ON THE INTERIOR OF THE EARTH DERIVED FROM SEISMIC SOURCES

JAMES B. MACELWANE, S. J.

(Professor of Geophysics and Director of the Department of Geophysics, Saint Louis University)

It is well known[16,51,62,64,128,139]* that the earth is an elastic medium and that a disturbance set up in an isotropic elastic solid will, in general, generate two types of elastic body waves. The first of these is a condensation-rarefaction wave which involves change of volume. The second type is a shear wave in which there is only distortion without change of volume. We shall speak of them as *condensation waves* and *shear waves*. Let S be the vector displacement, and let **u**, **v** and **w** be the rectangular components of S parallel, respectively, to the right-handed Cartesian axes, x, y and z. Let θ be the cubical dilatation which is equal to the divergence of the displacement.

$$\text{Div } \mathrm{S} = \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = \theta. \tag{19}$$

Let ω be the vector angular velocity of distortion and let ω_x, ω_y and ω_z be the respective components of this angular velocity about the axes x, y and z. Then, if **i**, **j** and **k** are parallel to x, y and z,

$$\text{curl } \mathrm{S} = 2\omega. \tag{20}$$

$$2\omega_x = \left(\frac{\partial w}{\partial y} - \frac{\partial v}{\partial z}\right)\mathbf{i}, \qquad 2\omega_y = \left(\frac{\partial u}{\partial z} - \frac{\partial w}{\partial x}\right)\mathbf{j}, \qquad 2\omega_z = \left(\frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}\right)\mathbf{k}. \tag{21}$$

Let E be Young's modulus, μ the modulus of rigidity, σ Poisson's ratio and k the bulk modulus or modulus of incompressibility, all defined in the usual way. It will be convenient also to define Lamé's compression constant λ. It is related to Young's modulus and Poisson's ratio by the equation

$$\lambda = \frac{\sigma E}{(1 + \sigma)(1 - 2\sigma)}. \tag{22}$$

* Numbers refer to bibliography at the end of this chapter.

The following relations are then known to hold:

$$\mu = \frac{E}{2(1 + \sigma)}. \tag{23}$$

$$E = \frac{\mu}{\lambda + \mu}(3\lambda + 2\mu) = \frac{\text{linear stress}}{\text{linear strain}}. \tag{24}$$

$$\sigma = \frac{\lambda}{3(\lambda + \mu)} \tag{25}$$

$$k = \frac{E}{2(1 - 2\sigma)} = \lambda + \tfrac{2}{3}\mu. \tag{26}$$

$$\lambda = k - \tfrac{2}{3}\mu. \tag{27}$$

The vector equation of small motion of an isotropic elastic solid may now be written thus:

$$\rho\frac{\partial^2 S}{\partial t^2} = (\lambda + \mu)\nabla\theta + \mu\nabla^2 S. \tag{28}$$

If we take the divergence of this equation of motion, we obtain the following equation:

$$\frac{\partial^2\theta}{\partial t^2} = \frac{\lambda + 2\mu}{\rho}\nabla^2\theta. \tag{29}$$

This is evidently the equation of a wave in which the cubical dilatation θ occurs as the variable. It is, therefore, a condensation wave whose velocity is given by the square root of the constant coefficient. Calling the velocity of this condensation wave V_c, we have the equation

$$V_c^2 = \frac{\lambda + 2\mu}{\rho} = \frac{k + \frac{4}{3}\mu}{\rho}. \tag{30}$$

Taking the curl of the equation of motion and substituting, we obtain the three component equations

$$\begin{aligned} \frac{\partial^2\omega_x}{\partial t^2} &= \frac{\mu}{\rho}\nabla^2\omega_x; \\ \frac{\partial^2\omega_y}{\partial t^2} &= \frac{\mu}{\rho}\nabla^2\omega_y; \\ \frac{\partial^2\omega_z}{\partial t^2} &= \frac{\mu}{\rho}\nabla^2\omega_z. \end{aligned} \tag{31}$$

These three equations represent the rectangular components of wave motion in which the variable is the curl of the displacement. They, therefore, represent a shear wave which is propagated with a

velocity equal to the square root of the constant. Calling this velocity V_s, we have

$$V_s^2 = \frac{\mu}{\rho}. \tag{32}$$

Substituting the values of V_c and V_s, we have the further relations

$$\left(\frac{V_c}{V_s}\right)^2 = 2\left(\frac{1-\sigma}{1 - 2\sigma}\right)$$

$$k = \rho(V_c^2 - \tfrac{4}{3}V_s^2). \tag{34}$$

$$\lambda = \rho(V_c^2 - 2V_s^2). \tag{35}$$

$$\mu = \rho V_s^2. \tag{36}$$

$$\sigma = \frac{V_c^2 - 2V_s^2}{\;}$$

Therefore, we see that it is possible to determine the elastic constants of an isotropic elastic solid provided that we are able to measure independently the velocities of the two types of elastic body waves, V_c and V_s and the density ρ. However, the earth is not isotropic; so that the simple theory of elasticity for an isotropic solid cannot be applied to the earth directly without further consideration. In the general case of an elastic solid, 36 constants are required to determine the motion. Each degree of symmetry introduces a relationship between some of these constants and permits elimination. In the limiting case of an isotropic solid, the 36 constants are reduced to 2, that is E and σ, or λ and μ, or k and μ. Hence, in the case of our partly aeolotropic earth, it is necessary to determine what degrees of symmetry, at least on a large scale, are present. There is no direct means of attack on this problem; but we know that the accessible portion of the earth's crust is made up of layers of rocks, samples of which may be studied in the laboratory. Many of these layers are reasonably homogeneous, and the specimens taken from some of them are approximately isotropic in the gross. Measurements of the elastic properties of rocks have been made by a number of investigators.[1,2,3,4,14,15,17,18,112,113,114,158,160] In some types of rock there is relatively detailed homogeneity within the layers and isotropic symmetry in the individual specimens, even when the constituent grains are crystalline in character and therefore aeolotropic in themselves and diversely oriented. But the departures from completely isotropic symmetry in other cases are considerable. In the aggregate the outer crust of the earth is so heterogeneous that we are obliged to determine the characteristics of each individual layer. At greater depths the earth seems to be constituted of material whose

properties vary much less radically. Usually the variation is continuous, but at certain definite horizons it would seem to be discontinuous. We shall consider the seismic evidence for the properties (1) of the outer layers near the surface, (2) of the deeper crust, (3) of the deeper mantle and intermediate material and (4) of the core of the earth.

SEISMIC DATA IN REGARD TO THE OUTER SURFACE LAYERS

Much information on the structure of the outermost layers of the earth's crust has been secured by the refraction and reflection methods of seismic prospecting. Although the results of such studies as have been made under commercial auspices are not fully available and the measurements of seismic wave velocities that have been published are not all of equal value, the results included in Table 25 with references to the respective publications in which they have appeared will give the critical reader an opportunity to form his own judgment concerning the velocities characteristic of certain rocks. Tables 26 and 27 show the tendency for velocities in similar rocks to increase with geologic age and with depth of burial.

In preparation for measuring the depth of the Greenland icecap by the seismic method a number of studies of glacial thickness were made in Europe. Mothes[153] chose for this purpose the Hintereisferner in the Austrian Alps because its profile was known from measurements made by Hess.[80] From a large number of records Mothes obtained an average velocity of the condensation waves of 3.14 km. per second in the névé and 3.6 km. per second in the ice of the tongue of the glacier and a velocity of the shear waves of 1.35 km. per second in the névé and 1.69 km. per second in the tongue. Another series of measurements were made by Mothes[154] on the Great Aletsch Glacier in March, 1929. He found the average velocity of the condensation waves in the ice of the Konkordia Platz to be 3.57 km. per second. In the summer of 1929, Brockamp and Mothes[20] made a series of seismic measurements on the Pasterzegletscher in the eastern Alps. They found a series of waves as follows: (1) a condensation wave with a velocity of 3.58 km. per second; (2) a shear wave with a velocity of 1.67 km. per second; (3) a condensation wave which was refracted into the rock underneath the glacier and traveled through it with a velocity of 5.85 km. per second; (4) a conducted condensation wave which they interpreted as having followed the boundary between the ice and rock with the speed of condensation waves in the ice; (5) shear waves which they likewise interpreted as having followed the lower boundary of the glacier with the speed of shear waves in ice; (6) reflected condensation

TABLE 25

OBSERVED VELOCITIES OF CONDENSATION WAVES IN VARIOUS MEDIUMS

Material	Locality	Velocity, km./sec.	Authority
Alluvium	Owens Valley	0.9 –1.0	Gutenberg, Wood and Buwalda[74]
Anhydrite	U. S. mid-continent and Gulf coast	4.1	Pirson[163]
Chalk:	U. S. mid-continent and Gulf coast	3.7	Pirson[163]
Austin	Texas	3.6 –4.2	Barton[10]
Cenon	France	2.1	Maurin and Éblé[144]
Cretaceous	North Germany	2.1 –2.3	Barsch and Reich[9]
Pecan Gap	Texas	3.0 –3.6	Barton[10]
Clay:	U.S.S.R.	2.1	Kirnos, Koridalin, Masarsky and Raiko[108]
Liassic	Göttingen	2.5	Ramspeck[166]
Cemented sandy	New South Wales	1.2 –1.3	Edge and Laby[47]
Miocene arenaceous	North Germany	1.6 –1.7	Barsch and Reich[9]
Recent or Tertiary	?	1.25–3.0	Gutenberg[68]
Surface	Göttingen	1.1	Müller[157]
Second layer	Göttingen	2.4	Müller[157]
Glacial	Alberta, Can.	1.4 –1.7	Heiland[73]
Clays and marls, Pleistocene	North Germany	1.6	Barsch and Reich[9]
Crystallines:			
Basement	Atlantic coastal plain, Virginia	5.2 –6.1	Ewing, Crary and Rutherford[49]
Old	Frazier Mountain	4.0 –4.25	Gutenberg[68]
Old	Beartooth Mountains	*ca.* 5.5	Gutenberg[65]
Glacial drift, sandy, dry	Alberta, Can.	0.4 –0.5	Heiland[73]
Glacial drift, sandy, wet	Alberta, Can.	0.9 –1.2	Heiland[73]
Gneiss (?)	Pennsylvania	5.39	Ewing and Crary[48]
Gneiss and schist	Alabama Hills, Owens Valley, Calif.	3.1	Gutenberg, Wood and Buwalda[74]
Granite:	New South Wales	5.6	Edge and Laby[47]
	Yosemite	5.25	Gutenberg, Wood and Buwalda[74], Gutenberg[68]
Quincy	Massachusetts	5.0	Leet and Ewing[123]
Rockport	Massachusetts	5.1	Leet and Ewing[123], Leet[34]
Westerly	Massachusetts	5.0	Leet and Ewing[123]
Granodiorite	New South Wales	4.6	Edge and Laby[47]
Gravel and sand:			
Dry	California, Wyoming	0.5 –1.0	Gutenberg[68]
Pleistocene	Sperenberg	0.9	Schweydar and Reich[179]
Gypsum	Jena	2.0	Meisser[146]
	Sperenberg	3.5	Schweydar and Reich[179]
	U. S. mid-continent and Gulf coast	3.0	Pirson[163]
Gypsum and red beds:			
Triassic	Big Horn basin, Wyo.	2.75–3.0	Gutenberg[65]
Gypsum and sandstone	Kunitz	2.5	Meisser[146]
Hornfels	New South Wales	3.5 –4.4	Edge and Laby[47]
Ice:			
Glacier	Alps	3.6	Mothes[153,154]
			Brockamp and Mothes[2]
Glacier	Greenland	3.5	Sorge and Loewe[159]
Lake	Lake in Germany	3.2	Köhler[110]
		2.7	Köhler[110]
		2.2	Köhler[110]
Lake	Saylor's Lake, Monroe Co., Pa.	3.46	Ewing, Crary and Thorne
Canal	Lehigh Canal, Pa.	3.28	Ewing, Crary and Thorne
Barrier	Ross Sea, Antarctica	3.7	Second Byrd expedition
Limestone:	U. S. mid-continent and Gulf coast	.4 –6.1	Pirson[163]
Dolomitic	Pennsylvania	5.97	Ewing and Crary[48]
Edwards	U. S. mid-continent and Gulf coast	3.4	Pirson[163]
Leesport cement rock	Pennsylvania	7.07	Ewing and Crary[48]
Schaumkalk	Rudersdorf	4.3	Schweydar and Reich[179]
Very hard	?	5.4	Rutherford[173]
Viola	Oklahoma	5.42	Rutherford[174]
	U. S. mid-continent and Gulf coast	3.9	Pirson[163]
Loam:	New South Wales	0.8	Edge and Laby[47]
Sandy	Jena	0.3	Meisser[146]
Loam and marl	North Germany	1.6	Barsch and Reich[9]

TABLE 25.—(*Continued*)

Material	Locality	Velocity, km./sec.	Authority
Loess	Göttingen	0.6	Ramspeck[1?]
	Kahla, Thuringia	0.3	Meisser[146]
Névé	Alps	3.1	Mothes[153]
Norite	Sudbury	6.2	Leet[118]
Salt:			
Rock	U. S. mid-continent and Gulf coast	4.6	Pirson[163]
of Salt domes	Texas-Louisiana	4.7 –5.2	Barton[10]
Sand:			
Calcareous	Kahla, Thuringia	0.8	Meisser[146]
Cemented	New South Wales	0.9 –1.0	Edge and Laby[47]
Oligocene, argillaceous	North Germany	1.6 –1.7	Barsch and Reich[9]
Pleistocene, dry	North Germany	0.7 –1.0	Barsch and Reich[9]
Unconsolidated	U. S. mid-continent and Gulf coast	0.9 –1.8	Pirson[163]
Wet	Kummersdorf	1.0	Schweydar and Reich[179]
Wet	Germany	1.4	Hecker[75]
Wet	California	0.75–1.5	Gutenberg[68]
Sandstone:			
Calcareous	U. S. mid-continent and Gulf coast	3.0 –4.0	Pirson[163]
Cretaceous	Wyoming	2.0 –2.5	Gutenberg[68]
Dakota	Yellow Cat Dome, Utah	3.00	Smith and Wilson[188]
Siliceous	U. S. mid-continent and Gulf coast	2.4– 3.4	Pirson[163]
Sandstone conglomerate	New South Wales	2.4	Edge and Laby[47]
Sediments:			
Carboniferous	Villaneuva de las Minas, Spain	3.1 –3.7	Siñeriz[186]
Eocene, middle	Texas-Louisiana	4.0	Barton[10]
Miocene, upper	Texas-Louisiana Gulf Coast	2.4 –2.7	Barton[10]
Oligocene	Mississippi	3.8 –4.3	Barton[10]
Pleistocene	Lüneberger Heide	1.6	von Schmidt[175]
Pliocene-Pleistocene	Texas Gulf coast	2.0	Barton[10]
Upper zone, Cretaceous to Recent	Atlantic coastal plain, Virginia	1.5 –1.8	Ewing, Crary and Rutherford[49]
Lower zone, Triassic-Jurassic?	Atlantic coastal plain, Virginia	2.0 –2.6	Ewing, Crary and Rutherford[49]
Slate, hard clay	New South Wales	3.2 –3.5	Edge and Laby[47]
Mancos	West of Cisco, Utah	4.12	Smith and Wilson[188]
Morrison formation, Cretaceous or Jurassic	Wild Cat Dome, Utah	2.71	Smith and Wilson[188]
Paradox, Pennsylvanian	Salt Valley anticline, Utah	3.01	Smith and Wilson[188]
Shale and quartzite:			
Carboniferous	North Germany	5.0	Barsch and Reich[9]
Shale and sandstone:			
Carboniferous	North Germany	3.8	Barsch and Reich[9]
Slate, hard clay	New South Wales	3.2 –3.5	Edge and Laby[47]
Syenite, nepheline	Saline Co., Ark.	5.5	Leet and Ewing[122]
"Weathered" layer:	U. S. mid-continent and Gulf coast	0.3 –0.9	Pirson[163]
	U. S. mid-continent and Gulf coast	0.6 –0.8	Lester[124]
Upper	Colorado	0.4	Heiland[77]
Lower	Colorado	1.12	Heiland[77]

waves. Since these waves penetrated to a considerable depth so that the elastic constants and hence the velocities were altered by pressure, Köhler[110] undertook an investigation of the speed of elastic waves in lake ice. He found that near the shore three presumably condensation waves were observed with a velocity of 3.2 km. per second, 2.7 km. per second and 2.2 km. per second, respectively, and one shear wave with a velocity of 1.45 km. per second, whereas near the center of the lake,

TABLE 26

VARIATION OF VELOCITY OF CONDENSATION WAVES WITH DEPTH

(According to Gutenberg[68])

Locality	Limiting depth of refraction profile	Velocity, km./sec.
Wyoming:		
Big Horn basin	Triassic red beds near surface	2.4
	At 2,000 m.	4.5
California:		
Los Angeles basin	Surface below thin low-velocity layer	1.9
	At 2,000 m.	3.5
San Joaquin Valley	Surface below thin low-velocity layer	<2
	At 2,000 m.	ca. 3
Yosemite	Granite near surface	5.25
	At 2,000 m.	5.5 ±

where the ice was probably isotropic, one condensation wave with a velocity of 3.2 km. per second was observed and one shear wave with a velocity of 1.7 km. per second.

Ewing, Crary and Thorne[50] also measured the velocity of propagation of elastic waves in the ice of Saylor's Lake and on the Lehigh Canal in Pennsylvania. The values they obtained for the speed of condensation waves were 3.46 and 3.28 km. per second, respectively.

Sorge[189] and Loewe accompanied the Wegener expedition to Greenland in 1929. The seismic measurements on the depth of the inland ice showed, at greater distances, velocities of 3.47 km. per second for the condensation waves and 1.7 km. per second for the shear waves which traversed the ice into which the snow fields gradually passed with depth, thus giving the value $\sigma = 0.342$ for Poisson's ratio. The velocities increased with further depth. The results for condensation waves in ice are summarized in Table 25.

Wiechert[211,212] carried on his observations of velocities in surface rocks during the years from 1906 to 1929. In the neighborhood of Göttingen, three paths for the transmission of condensation waves were found P_1, P_2 and P_3. This observational material was gathered together and published after his death by his students Brockamp and Wölcken,[19,21] who added some observations of their own.

Brockamp[19,21] and Wölcken undertook a further study of the velocities of the condensation waves from the quarry blasts that had been observed by the geophysical staff of the University of Göttingen. They reduced all shot points and observing stations to a common

TABLE 27

VARIATION OF THE VELOCITY OF CONDENSATION WAVES WITH DEPTH AND GEOLOGICAL AGE

(Data drawn from Wells in the states of Colorado, Kansas, Louisiana, Mississippi, New Mexico, Oklahoma, Pennsylvania, and Texas after Weatherby and Faust[208])

Geological age	Velocities in shale and sandstone, km./sec. At depth of 0–600 m.	At depth of 600–900 m.	At depth of 900–1,200 m.
Devonian...........	4.1	4.2	4.2
Pennsylvanian.......	2.9	3.4	3.5
Permian............	2.6	3.0	
Cretaceous..........	2.3	2.8	3.3
Eocene.............	2.2	2.7	3.1
Oligocene-Pleistocene	2.0	2.2	2.5

Velocities in limestone

Geological age	Formation name	Average depth of column measured	Velocity, km./sec.
Cambro-Ordovician.	Arbuckle	At the surface	5.3
Ordovician.........	Viola	At the surface	5.1
		110 m.	6.1
Devonian.....	Hunton	At the surface	4.3
		140 m.	5.3
Mississippian..	Mayes	At the surface	3.8
		140 m.	5.2
Pennsylvanian.	Belle City	At the surface	4.6
		90 m.	4.7
Permian......		110 m.	4.7
Cretaceous....	Edwards	At the surface	3.4
		100 m.	4.1

elevation of 270 m. above mean sea level by introducing in the observed travel times a correction corresponding to the difference between the actual and the reference levels divided by the observed velocity of condensation waves in the respective surface rocks. Their values for P_1 are given in Table 28.

P_1 was found to have widely different velocities in the various localities, as was to be expected. Its velocity in the basalt of the Vogelsberg was 5.6 km. per second up to a distance of 12 km. from the

TABLE 28

VELOCITIES OF CONDENSATION WAVES IN VARIOUS TYPES OF ROCK

[After Brockamp (Ref. 19, p. 296)]

Rock	Velocity, km./sec.	Authority
Basalt	5.6	Brockamp-Wölcken[19,21]
Limestone	4.2	Schweydar-Reich[176]
Mottled sandstone	2.2	Meisser-Martin[145]
Tertiary sands and clays	1.7	Brockamp-Wölcken[19,21]
Basement rocks	5.5	Brockamp-Wölcken[19,21]
Alternating chalk and limestone	3.5	Brockamp-Wölcken[19,21]

shot point. In the neighborhood of Göttingen the velocity of P_1 was 3.5 km. per second to a distance of 16 km. In the neighborhood of Treysa in the Hessian trough up to a distance of 10 km. it had a velocity of only 1.7 km. per second.

The P_2 phase was found to have a linear time-distance graph whose slope corresponded to an average velocity of 5.9 km. per second, thus seeming in the opinion of Brockamp[19] to identify the P_2 of Wiechert[212] with the $\bar{P}$ of Mohorovičić[149] and the P_g of Jeffreys[91] in the records of near earthquakes. However, the velocities found for $\bar{P}$ are usually somewhat smaller, as will be seen presently.

Other values for the velocities of elastic waves which have been determined from the records of explosions are given in Tables 29 and 30.

TABLE 29

VELOCITIES OF CONDENSATION WAVES IN VARIOUS TYPES OF ROCK

(After Thoenen and Windes)

Rock type	Observed velocities, km./sec.		
	Minimum	Average	Maximum
Biotite-gneiss	...	8.0	
Dolomite	...	6.9	
Flint	3.8	4.6	4.9
Gabbro-diorite	4.9	5.4	5.7
Limestone	2.4	4.0	5.8

Gutenberg[69] and Richter think that the P_2 of Wiechert[212] and Brockamp[19] does not correspond to $\bar{P}$ but had traversed sediments in which the velocity is higher than that in granite and that the layer which elsewhere is responsible for $\bar{P}$ underlies these sediments in the neighborhood of Göttingen.

TABLE 30

VELOCITIES OF WAVES CAUSED BY EXPLOSIONS AND BLASTS

Location	Wave type	Velocity, km./sec.	Authority
Europe			
France:			
La Courtine	P	4.9	Maurain, Éblé and Labrouste[145]
	P	5.3	Maurain, Éblé and Labrouste[145]
	P	5.5	Maurain, Éblé and Labrouste[145]
	P	5.6	Maurain, Éblé and Labrouste[145]
	P	6.2	Maurain, Éblé and Labrouste[145]
	P	5.5	Rothé, Lacoste, Bois, Dammann and Hée[171,145]
Germany:			
Hainberg	P	3.36	Müller[156]
Oppau	P_g	5.53	Jeffreys[103]
	P	5.4	Wrinch and Jeffreys[219]
	P	5.73	Hecker[76]
	P	5.4–5.6	Gutenberg[59]
Italy:			
Carrara	P	4.6	Agamennone[6]
	S	3.0	Agamennone[6]
Falconara	P	6.2–6.4	Quervain[165]
	S	3.64	Quervain[165]
Switzerland:			
Alpnach	P	4.7	Quervain[165]
Grenchen	P	5.1–5.25	Quervain[165]
North America			
United States:			
California			
Los Angeles basin	P	2.9–3.5	Gutenberg, Wood and Buwalda[74]
Richmond	P_1	4.3	Byerly and Wilson[32]
	P_2	5.4	Byerly and Wilson[32]
	S_1	2.4	Byerly and Wilson[32]
	S_2	3.1	Byerly and Wilson[32]
	S_3	3.8	Byerly and Wilson[32]
San Gabriel	P	5.5	Wood and Richter[216]
Southern California	P	4.1	Wood and Richter[218]
	P	5.0	Wood and Richter[218]
	P	5.4	Wood and Richter[218]
	P	5.55	Wood and Richter[218]
	P	5.9	Wood and Richter[218]
	P	6.0	Wood and Richter[218]
	S	2.7	Wood and Richter[218]
	S	3.0	Wood and Richter[218]
	S	3.15	Wood and Richter[218]
	S	3.21	Wood and Richter[218]
	S	3.25	Wood and Richter[218]
	S	3.4	Wood and Richter[218]
	S	3.5	Wood and Richter[218]
Ventura basin	P	2.9–3.5	Gutenberg, Wood and Buwalda[74]
Victorville	P	5.5	Wood and Richter[218]
New England	P	6.0	Leet[119]
	P	8.0	Leet[119]
	S	3.5	Leet[119]
	S	4.6	Leet[119]
	P_1	6.01	Leet[121]
	P_2	6.77	Leet[121]
	S_1	3.45	Leet[121]
	S_2	3.93	Leet[121]

The phase P_3 of Wiechert[212] and of Brockamp[19] was found to have a velocity of 6.72 km. per second. Brockamp is of the opinion that this phase corresponds to the P^* of Conrad.[36] Gutenberg's P_m[66] in southern California has a similar velocity.

Conrad[37] found in his investigation of the Schwadorf earthquake that a distinct condensation wave arrived ahead of the waves just mentioned and at distances over 200 km. became quite prominent. He found it to have a velocity of 7.87 km. per second. Gutenberg[66] found a similar wave in the southern California earthquakes with a velocity of 7.60 km. per second.

In the records of near earthquakes many phases have been identified by working seismologists. However, there is much disagreement as to the interpretation to be put upon them. What changes of amplitude or period or both are sufficiently conspicuous to be characterized as new phases in a record already disturbed? All would agree that a phase is real if it can be followed from station to station in such wise that the respective arrival times when plotted against epicentral distances are isolated and lie on or very close to a continuous curve. But there is frequently so much scattering of the points and so many other points near by that the choice of a particular curve would seem arbitrary. Even in the case of first arrivals, two doubts arise. (1) Was the seismograph sufficiently sensitive to pick up a small impulse that was actually the first to arrive at the station? (2) Did a small impulse that was first recorded at nearer stations persist until it arrived at the station in question, or did it suffer extinction on the way? There seems to be evidence that small impulses do persist and can be recorded at surprising distances by seismographs that are sufficiently sensitive. Therefore, difficulties with first arrivals in near earthquakes should be reducible either to inadequate recording or to the presence of microseisms.

It is thought by Jeffreys[102] and others that the crust of the earth is, in general, two-layered and hence that there are three main paths for impulses or waves in near earthquakes: $\bar{P}$ or P_g and $\bar{S}$ or S_g directly through the upper layer from focus to station; P^* and S^* refracted into the second layer at the critical angle or diffracted by it and traveling along the interface with the speed characteristic of the lower medium and leaving it continuously at the critical angle to return to the surface; and lastly P_n and S_n entering and leaving the substratum beneath the crust under similar conditions and traveling along the base of the crust with the speed characteristic of the substratum. Schmidt[176,177,178] developed an interesting theory to account for the great energy observed in the refracted impulses in seismic prospecting. His theory

TABLE 31

VELOCITIES OF $\bar{P}$ (OR P_g)

Locality	Earthquakes	Velocity, km./sec.	Authority
Asia	Central Asian	5.54	Rozova[172]
Japan	Japanese	5.0	Matuzawa[141]
			Matuzawa, Yamada and Suzuki[143]
	Mount Asama (volcanic, Sept. 18, 1929)	3.56	Isikawa[90]
Europe:			
Austria	North Tyrol	5.7	Gräfe[54]
		5.724	Jeffreys[97]
	Schwadorf	5.60	Conrad[37]
		5.598	Jeffreys[97]
	Tauern	5.4	Conrad[36]
		5.627	Jeffreys[97]
Belgium	North Brabant	5.62	Gees[52]
Central and western	....	5.570	Jeffreys[103]
England	Herefordshire	5.652	Jeffreys[97,103]
France:			
Orne	Briouze-St. Gervais	5.4	Mourant[155]
North of Brittany coast	English Channel	5.4	Mourant[155]
West coast of Normandy	Jersey	5.4	Mourant[155]
		5.441	Jeffreys[97,103]
Germany:			
Baden	Lake Constance	5.55	Hiller[82]
Prussia	Rhineland	5.6–6.0	Gutenberg[61]
	Saar	5.6	Landsberg[115]
Württemberg	South German I	5.6–6.0	Gutenberg[56]
		5.556	Jeffreys[97]
	South German II	5.6–6.0	Gutenberg[56]
		5.522	Jeffreys[97]
Greece:			
Cephalonia	Argostolion	5.8	Stoneley[195]
Italy	Carnic Alps	5.7	Caloi[33]
Switzerland	Visp	5.70	Wanner[207]
		5.57	Quervain[164]
	Yverdon	5.75	Wanner[207]
Yugoslavia:			
Croatia	Kulpa Valley	5.6	A. Mohorovičić[149]
		5.637	Jeffreys[97,103]
Dalmatia	Imotski	5.50	Tillotson[199]
New Zealand	Gisborne-Wairoa	5.5	Bullen[25]
North America:			
United States:			
California	Niles	5.4	Byerly and Wilson[31]
	Parkfield	5.6	Byerly and Wilson[31]
	Sierra Nevada	5.5	Byerly[26]
	21 earthquakes in southern California	5.55	Gutenberg[66]
	Whittier	5.55	Wood and Richter[217]

TABLE 32
VELOCITIES OF P^*

Locality	Earthquakes	Velocity, km. sec.	Authority
Asia	Central Asian	5.99	Rozova[172]
Japan	Japanese	6.1	Matuzawa[141]
		6.2	Matuzawa, Yamada and Susuki[143]
	Tango	6.3	Hodgson[84,86]
Europe:			
Austria	North Tyrol	6.7	Gräfe[54]
		7.082	Jeffreys[97]
	Schwadorf	6.47	Conrad[57]
		6.468	Jeffreys[97]
	Tauern	6.29	Conrad[56]
		6.254	Jeffreys[97]
Belgium	North Brabant	6.42	Gees[52]
Central and western	....	6.498	Jeffreys[96]
England	Herefordshire	6.3	Jeffreys[92]
France:			
Orne	Briouze-St. Gervais	6.3	Mourant[155]
North of Brittany coast	English Channel	6.3	Mourant[155]
West coast of Normandy	Jersey	6.3	Jeffreys[92] and Mourant[155]
Germany:			
Baden	Lake Constance	6.3	Hiller[82]
Württemberg	South German		
	I	7.1	Gutenberg[66]
	I	6.30	Jeffreys[97]
	Mean	6.5	Gutenberg[66]
Greece:			
Cephalonia	Argostolion	6.1	Stoneley[195]
Italy	Carnic Alps	6.4	Caloi[33]
Yugoslavia:			
Dalmatia	Imotski	6.30	Tillotson[199]
New Zealand	Gisborne-Wairoa	6.3	Bullen[25]
North America:			
United States:			
California	Sierra Nevada	7.4	Byerly[26]
Hawaii	....	7.2	Jones[105,106]

would seem to be applicable to the P^*, S^* and P_n, S_n phases of near earthquakes. They would be called *traveling reflections.*

The velocities of $\bar{P}$ or P_g, P^* and P_n that have been observed in various earthquakes are given in Tables 31, 32 and 33, respectively. The velocities of other phases of the P-type will be found in Table 37.

TABLE 33
Velocities of P_n

Locality	Earthquakes	Velocity, km./sec.	Authority
Asia:			
Central	Central Asian	7.82	Rozova[172]
Japan	Hatidyo Islands	8.48 (200 km. depth)	Honda[88]
	Japanese	7.5	Matuzawa[141]
			Matuzawa, Yamada and Suzuki[143]
	Lake Tazawa	7.7 (130 km. depth)	Oka[161]
	Tango	7.75	Hodgson[84,86]
Europe:			
Austria	North Tyrol	8.3	Gräfe[54]
		8.23	Jeffreys[97]
	Schwadorf	8.12	Conrad[37]
		8.104	Jeffreys[97]
	Tauern	7.83	Conrad[36]
		7.65	Jeffreys[97]
Belgium	North Brabant	7.63	Gees[52]
Central and western	....	7.764	Jeffreys[95]
England	Dogger Bank	7.86–8.43	Gees[52]
	Herefordshire	7.8	Jeffreys[92]
France:			
Orne	Briouze-St. Gervais	7.8	Mourant[155]
North of Brittany coast	English Channel	7.8	Mourant[155]
West coast of Normandy	Jersey	7.8	Jeffreys[91] and Mourant[155]
Germany:			
Prussia	Saar	8.05	Landsberg[115]
Württemberg	South German I	7.6 –8 0	Gutenberg[56]
		7.75	Jeffreys[97]
	South German II	8.2	Gutenberg[66]
		8.11	Jeffreys[97]
Greece:			
Cephalonia	Argostolion	7.8	Stoneley[195]
Ionian Islands	Ionian Island	7.68	Agamennone[5]
Italy	Adriatic	7.77	Caloi[34]
	Carnic Alps	7.81	Caloi[33]
Norway	North Sea	7.82	Lee[116]
Switzerland	Visp	7.70	Wanner[207]
Yugoslavia:			
Croatia	Kulpa Valley	7.9	A. Mohorovičić[149]
Dalmatia	Imotski	7.80	Tillotson[198]
New Zealand	Gisborne-Wairoa	8.10	Bullen[25]
North America:			
United States:			
California	Eureka	7.8	Sparks[190]
	Niles	7.9	Byerly and Wilson
	Northern coast	7.78, 7.83, 7.84	Byerly[29]
	Parkfield	8.0, 8.3	Byerly and Wilson
	Sierra Nevada	8.6	Byerly[26]
	Southern California	7.94	Gutenberg[66]
Nevada	Cedar Mountain	8.27	Byerly[28]
New England		8	Leet[120]
Territory of Hawaii	Hawaii	8.0	Jones[105,106]
Texas	Van Horn	8.0	Byerly[27]

VELOCITIES OF SHEAR WAVES IN THE SURFACE LAYERS

Very prominent shear waves were observed on the records of blasts in stone quarries about Göttingen which were investigated by Wiechert[211,212] and by Brockamp[19,21] and Wölcken. From a detailed study of these impulses, Korte[111] found that the time-distance graph could not be distinguished from a straight line in the range 7 to 230 km. and that the corresponding velocity was 3.4 km. per second. The ratio of the P_2 velocity (5.9 km. per second) to this velocity is 1.735

TABLE 34

VELOCITIES OF $\bar{S}$ (OR S_g)

Locality	Earthquakes	Velocity, km./sec.	Authority
Asia:			
Central	Central Asian	3.29	Rozova[172]
Japan	Japanese	3.15	Matuzawa[141]
			Matuzawa, Yamada and Suzuki[143]
	Mount Asama (volcanic)	2.22 (at surface)	Isikawa[90]
Europe:			
Austria	North Tyrol	3.5	Gräfe[54]
		3.466	Jeffreys[97]
	Schwadorf	3.39	Conrad[37]
		3.41	Jeffreys[97]
	Tauern	3.370	Jeffreys[97]
Belgium	North Brabant	3.46	Gees[52]
Central and western	....	3.363	Jeffreys[95]
England	Herefordshire	3.372	Jeffreys[97,103]
France:			
Orne	Briouze-St. Gervais	3.3	Mourant[155]
North of Brittany	English Channel	3.3	Mourant[155]
West coast of Normandy	Jersey	3.3	Jeffreys[91] and Mourant[155]
		3.361	Jeffreys[97,103]
Germany:			
Baden	Lake Constance	3.31	Hiller[82]
Prussia	Saar	3.36	Landsberg[115]
Greece:			
Cephalonia	Argostolion	3.3	Stoneley[195]
Italy	Carnic Alps	3.3	Caloi[33]
Switzerland	Visp	3.34	Quervain[165]
		3.43	Wanner[207]
	Yverdon	3.47	Wanner[207]
Yugoslavia:			
Dalmatia	Imotski	3.30	Tillotson[199]
New Zealand	Gisborne-Wairoa	3.3	Bullen[25]
North America:			
United States:			
California	Niles	3.2	Byerly and Wilson[31]
	Parkfield	3.3	Byerly and Wilson[31]
	Sierra Nevada	3.3	Byerly and Wilson[31]
	Whittier	3.25	Wood and Richter[217]
Territory of Hawaii	Hawaii	3.3	Jones[106]

TABLE 35

VELOCITIES OF S^*

Locality	Earthquakes	Velocity, km./sec.	Authority
Asia:			
Central.	Central Asian	3.79	Rozova[172]
Japan.	Japanese	3.7	Matuzawa[141]
			Matuzawa, Yamada and Suzuki[143]
Europe:			
Austria.	North Tyrol	3.6	Gräfe[54]
		3.600	Jeffreys[97]
	Schwadorf	3.57	Conrad[37]
		3.604	Jeffreys[97]
	Tauern	3.57	Conrad[36]
		3.584	Jeffreys[97]
Belgium.	North Brabant	3.85	Gees[52]
Central and western. .	. . . .	3.741	Jeffreys[95]
England.	Herefordshire	3.7	Jeffreys[92]
France:			
West coast of Normandy.	Jersey	(S_1*)3.744	Jeffreys[97,103]
Germany:			
Baden.	Lake Constance	3.7	Hiller[82]
Greece:			
Cephalonia.	Argostolion	3.7	Stoneley[195]
Italy.	Carnic Alps	3.5	Caloi[33]
Yugoslavia:			
Dalmatia.	Imotski	3.65	Tillotson[199]
New Zealand:.	Gisborne-Wairoa	3.7	Bullen[25]
North America:			
United States:			
Territory of Hawaii	Hawaii	3.9	Jones[105,106]

and the corresponding value of Poisson's ratio is 0.255. Hence Korte concluded that these two phases are to be coordinated and that the velocity 3.4 km. per second is therefore the velocity of shear waves in the basement complex and the phase is to be designated S_2. The phase is very prominent on the horizontal component records and much less so on the vertical. Leet and Ewing[123] recorded a shear wave in the Quincy granite with a velocity of 2.48 km. per second. Compared with the corresponding condensation wave, it gives the abnormally high value of 0.333 for Poisson's ratio. In the Sudbury norite, Leet[118] found the velocity of shear waves to be 3.49 km. per second and the corresponding value of Poisson's ratio to be 0.27.

TABLE 36
VELOCITIES OF S_n

Locality	Earthquakes	Velocity, km. sec.	Authority
Asia:			
Japan	Hatidyo Islands	4.8 (200 km. depth)	Honda[83]
	Japanese	4.5	Matuzawa[142]
			Matuzawa, Yamada and Suzaki[143]
	Lake Tazawa	4.4 (130 km. depth)	Oka[161]
Europe:			
Austria	North Tyrol	4.4	Gräfe[54]
		4.446	Jeffreys[97]
	Schwadorf	4.383	Jeffreys[97]
Belgium	North Brabant	4.59	Gees[52]
Central and western	....	4.362	Jeffreys[96]
England	Dogger Bank	4.74–4.32	Gees[52]
	Herefordshire	4.436	Jeffreys[97]
France:			
Orne	Briouze-St. Gervais	4.35	Mourant[155]
North of Brittany	English Channel	4.35	Mourant[155]
West Coast of Normandy	Jersey	4.384	Jeffreys[97]
Germany:			
Prussia	Saar	4.5	Landsberg[115]
Greece:			
Cephalonia	Argostolion	4.4	Stoneley[195]
Italy	Adriatic	4.8	Caloi[34]
	Carnic Alps	4.2	Caloi[35]
Norway	North Sea	4.36	Lee[116]
Switzerland	Visp	4.50	Wanner[207]
Yugoslavia:			
Dalmatia	Imotski	4.35	Tillotson[199]
New Zealand	Gisborne-Wairoa	4.38	Bullen[25]
North America:			
United States:			
California	Northern coast	4.34	Byerly[29]
		4.35	Byerly[29]
	Parkfield	4.6	Byerly and Wilson[illegible]
	Sierra Nevada	4.6	Byerly[26]
	Southern California	4.45	Gutenberg[66]
New England	....	4.6	Leet[120]
Territory of Hawaii	Hawaiian	4.6	Jones[105, 106]

TABLE 37

VELOCITIES OF OTHER PHASES IN NEAR EARTHQUAKES

Locality	Earthquakes	Phase	Velocity, km./sec	Authority
Europe:				
Austria	Schwadorf	P_x	7.87	Conrad[37]
		P_x	7.852	Jeffreys[97]
		S_x	4.32	Conrad[37]
		d	4.19	Conrad[37]
		α	3.81	Conrad[37]
	Tauern	α	3.80	Conrad[36,37]
Belgium	North Brabant	P_n'	7.59	Gees[52]
		P_n''	7.72	Gees[52]
France:				
Orne	Briouze-St. Gervais	P_s	4.7	Mourant[155]
		S_s	3.13	Mourant[155]
North of Brittany	English Channel	P_s	4.7	Mourant[155]
		S_s	3.13	Mourant[155]
Germany:				
Prussia	Saar	P_x	7.56	Landsberg[115]
		b	5.34	Landsberg[115]
		d	4.12	Landsberg[115]
		α	3.93	Landsberg[115]
Greece:				
Cephalonia	Argostolion	P_s	5.0	Stoneley[195]
		S_s	3.1	Stoneley[195]
Italy	Carnic Alps	P_m?	6.9	Caloi[33]
		α	3.8	Caloi[33]
Yugoslavia:				
Dalmatia	Imotski	P_s	5.00	Tillotson[199]
		S_x	4.35	Tillotson[199]
		S_s	3.36	Tillotson[199]
North America:				
United States:				
California	Eureka	P	7.7	Byerly and Sparks[30]
		P'	7.7	Byerly and Sparks[30]
		P''	7.7	Byerly and Sparks[30]
		P	7.4	Byerly and Sparks[30] Sparks[190]
		P	7.0	Byerly and Sparks[30] Sparks[190]
	Niles	I	7.7	Byerly and Wilson[31]
		IV	6.7	Byerly and Wilson[31]
	Northern coast	F	4.35	Byerly[29]
	Parkfield	II	7.1–7.3	Byerly and Wilson[31]
		IV	6.6	Byerly and Wilson[31]
		V	5.0	Byerly and Wilson[31]
		VI	4.2	Byerly and Wilson[31]
	Sierra Nevada	$\bar{P}_s$	5.8	Byerly[26]
		S?	4.4, 3.4, 3.2	Byerly[26]
	Southern California	P_x	7.60	Gutenberg[66]
		a	7.10	Gutenberg[66]
		P_m	6.83	Gutenberg[66]
		b	6.56	Gutenberg[66]
		P_y	6.05	Gutenberg[66]
		c	5.94	Gutenberg[66]
		S_x	4.24	Gutenberg[66]
		S_y	3.66	Gutenberg[66]
		S_m	3.39	Gutenberg[66]
New England	....	P_6	6	Leet[120]
		$S_{3.5}$	3.5	Leet[120]
Territory of Hawaii	Hawaiian	P_6	5.2	Jones[105]
		P_3	3.6	Jones[105]
		P_2	3.2	Jones[105]
		P_1	2.65	Jones[105]
		S_6	2.9	Jones[105]
		S_3	2.0	Jones[105]
		S_2	1.8	Jones[105]
		S_1	1.36	Jones[105]

Values of the velocity of shear waves $\bar{S}$ or S_g, S^* and S_n in the crustal layers as observed in near earthquakes are given in Tables 34, 35 and 36, respectively. Those of other phases of the S-type are tabulated with those of the P-type in Table 37.

DISCUSSION OF THE ARGUMENT FOR CRUSTAL STRUCTURE FROM THE PHASES OF NEAR EARTHQUAKES

The assumption of layers separated by discontinuities in the earth's crust affords one explanation of the various sets of phases of *P*- and *S*-type which appear on the records of near earthquakes. The fact that the first arrival times when plotted against distance from the epicenter tend to lie on a broken curve, the successive segments of which, at least after the first and up to distances of several hundred kilometers, are straight lines, is a strong argument in favor of the hypothesis. So also is the continuation of these lines beyond their intersections when later arrivals are plotted. The reciprocals of the slopes of the successive straight lines are then the true velocities in the successive layers, since the time spent in traversing the overlying layers will be constant for all points on the given line and the increase of time with distance will be entirely due to travel under the upper boundary of the layer. Hence, it should be possible with a sufficient number of records of near-by stations to calculate the depth of the focus, or point from which the first vibrations originated, and also the depth of the successive layers. A number of such attempts have been made. A. Mohorovičić[149,150] assumed a single layer 60 km. in thickness to explain the two successive phases that he found in the Kulpa (or Kupa) Valley earthquake. But his data were reinterpreted by Jeffreys[91] and combined with those of Conrad and others to give a thickness of 12 km. for an upper layer, which he called *granitic*, and 25 km. for a second, intermediate layer, which he called *basaltic* [91, 93 (page 321)]. Jeffreys' studies of the Jersey and Herefordshire earthquakes[92] caused him to revise these depths to 10 km. and 20 km., respectively. Lee,[116] in his study of the North Sea earthquake and others, came to the conclusion that there is in northern Europe a sedimentary layer about 1 km. in thickness underlaid by a granitic layer 14 km. thick and a basaltic layer 15 km. thick. In the Balkans the thicknesses he found were 1 km., 11.5 km. and 22 to 33 km. Jeffreys[102] in 1937 subjected all the European data to a critical numerical test for consistency and found the uncertainties so great that he was finally forced to *assume* as consistent with the indications a thickness of 17 km. for the upper layer and 9 km. for the intermediate layer. He says: "The uncertainties that remain, even after combining the data from several near earthquakes and for surface waves, indicate the futility of attempts to get accurate estimates of the epicentre, the velocities, the thicknesses and the focal depth simultaneously from a single near earthquake." (Ref. 103, page 212). The present writer would add that the futility of trying to determine these quantities by combining heterogeneous

groups of observations of inadequate accuracy from many earthquakes is even greater. An accuracy of observation of the absolute time of arrival of each impulse to *at least the nearest tenth of a second* is required, and that degree of precision is nowhere available except in favorable instances in southern California. The lack of sufficient accuracy in the absolute values of the observed times of arrival of phases in near earthquakes has been noted by many workers.[25,38,84,121,170] This severe limitation must be kept in mind in evaluating all present and past attempts to determine crustal structure in any region.

In the neighborhood of the Tango Peninsula, Japan, Hodgson[84,86] found that the so-called continental or granitic layer is very thin or entirely absent and that the thickness of the so-called basaltic or intermediate layer is of the order of 16 km. Bullen[25] made a tentative solution for the Gisborne-Wairoa and other New Zealand earthquakes and found that the structure suggested for New Zealand was the following: Sedimentary layer 0.7 km. in thickness, granitic layer 0.3 km., intermediate layer 16.5 km. This is in agreement with Dahm's[39] tentative conclusion from the Hawke Bay earthquake that the New Zealand structure resembles that of the Tango region. Gutenberg[66] finds in southern California constants for the successive layers as given in Table 38. On the other hand, in the Pasadena records of a deep

TABLE 38

CRUSTAL STRUCTURE IN SOUTHERN CALIFORNIA

[After Gutenberg (Ref. 66, page 41)]

Depth of layer, km.	Mean depth of lower boundary, km.	Velocity of condensation waves, km./sec.	Velocity of shear waves, km./sec.	Ratio of velocities V_c/V_s	Poisson's constant σ
0–14	14	5.55	3.23	1.70	0.24
14–26	25	6.05	3.39	1.79	0.27
26–30	31	6.83	3.66	1.86	0.30
30–39	39	7.6	4.24	1.80	0.27
>39	..	7.94	4.45	1.78	0.27

earthquake Sharpe[185] found evidence for a surface layer 6.6 km. thick and a large discontinuity at a depth of 20 km. Byerly and Wilson[31] found tentatively a three-layered structure in central and northern California as follows:

Depth, km.	Speed, km./sec.
1–13	5.6
13–25	6.6
25–31	7.3
	8.0

In New England Leet[118] found from a large number of explosions and accurately timed quarry blasts that a surface layer exists which apparently is 23 km. in thickness and in which the velocity of condensation waves is 6 km. per second and that of shear waves 3.5 km. per second. Below this layer the respective velocities are 8 km. per second and 4.6 km. per second. For Missouri the data are as yet inadequate; but Miss Robertson,[170] following the method of Slichter and Sharpe, found that if the structure is two-layered the suggested thicknesses are 16 km. for the upper and 13 km. for the lower layer.

In other parts of the American continent the data regarding velocities are still insufficient for the calculation of probable values for the thicknesses and depths of layers.

There seems to be some evidence for the existence of a different type of structure under the Pacific Ocean. Angenheister[7,8] found that the smallest observed speeds both for condensation waves and for shear waves were considerably greater in the neighborhood of Apia, Samoa, than on the continent of Europe. But the means at his disposal did not make it possible to determine the velocities with any high degree of accuracy in the absence of other first-class stations in the neighborhood. Brunner[24,136] found that, in the case of the very deep Kermadec-Fiji earthquake of May 26, 1932, the travel times of condensation waves to an epicentral distance of 12° was 15 sec. less than half the travel time of a ray from a surface shock to 24° calculated on the basis of Hodgson's curve for the Tango earthquake, Mar. 7, 1927, and Dahm's velocities which are largely based on that curve, thus pointing toward a difference in structure between the central Pacific Ocean and Japan.

EVIDENCE FROM SURFACE WAVES

As the present writer (Ref. 139, Chap. 13) has pointed out elsewhere, the various types of surface waves that are observed in most shallow and normal earthquakes are not adequately represented by existing mathematical theory. Therefore it is premature to base conclusions on this theory. Furthermore, the difficulties of empirical interpretation of the data are not inconsiderable. Measurements of velocity depend on the identification of the same wave front or group front at successive distances from the epicenter. This might be attempted in four conceivable ways. (1) One might single out a large, persistent maximum amplitude and follow it from station to station as the present writer[129] did in the case of the California earthquake of Jan. 31, 1922; but if one does so one will find that the period of the wave lengthens in a complicated way as one progresses to greater distances and that the theory of simple sine-wave propagation and group

velocity is not applicable. (2) One might follow the main front of the entire surface phase. This is possible in many earthquakes. But one will find that the character of the waves in this front will not be the same in the records of one earthquake as it is in another, even from the same epicentral region, nor will it remain uniform at all distances in the records of the same earthquake. Besides, this method does not measure the velocities of the separate types of surface waves. (3) Several investigators have measured the travel times and the periods of large numbers of individual waves in the surface phases and have classified these statistically according to path and according to whether or not the wave form showed a vertical component. If no vertical component was present in the wave or its neighbors, the wave was classified as a *surface shear wave;* if a vertical component was prominent and a transverse component lacking, it was placed under the category of *Rayleigh waves* although the theory proposed by Lord Rayleigh is not adequate to describe them. They may be called *pseudo-Rayleigh waves.* The results of these investigations are very interesting but not simple. The proved phenomenon of *real increase of period* with travel precludes the assumption that waves of the same period arriving at different distances are identical. But there are two results of these observations that are very valuable and undoubtedly throw some light on crustal structure. It is found (1) that the speeds of waves of different final period over the same general path may be different and (2) that the speeds over paths of comparable length but across a different unit of the earth's surface are different. The results are presented in Tables 39 to 53.

TABLE 39

VELOCITIES OF SURFACE SHEAR WAVES ACROSS THE AMERICAS

(After Carder[35])

Epicentral region	Station	Arcual distance, °	Observed wave period, sec.	Observed mean speed, km./sec.
Alaska	Ottawa	38.6	60	4.44
	Rio de Janeiro	111	50	4.20
Baffin Bay	Mount Hamilton	43.5	25	3.68
California	Chicago	28	50	4.07
	Georgetown	34.8	17	3.53
	Ottawa	35.5	50	4.10
	Saskatoon	16.9	50	4.01
	Tucson	14.6	40	4.16
	Vieques	55.5	50+	4.06
	Washington, D. C	36.5	44	3.93
Central America	Berkeley	35.4	30	3.81
		37.5	30	3.61
		36	30	3.64

TABLE 39.—(*Continued*)

Epicentral region	Station	Arcual distance, °	Observed wave period, sec.	Observed mean speed, km./sec.
		35.8	30	3.70
				3.77
		39.2	30	3.75
		38.6	30	3.75
		41.2	35	3.91
		39.3	36	3.87
		37.1	36	3.80
		36.9	36	3.73
		37.5	40	3.74
		40.4	40	3.95
		43+	40+	3.74
		37.4	40	4.03
		35.8	40	3.81
		39.2	40	3.92
		41.2	40	3.98
		39.3	46	4.19
		43+	46	3.86
		37.4	48	4.07
		35.4	50	4.02
	Mount Hamilton	38.6	60	4.21
		36.9	38	3.91
		36.5	33	3.75
		36.8	30	3.75
		36.4	25	3.46
Chile	Rio de Janeiro	26.4	100	4.53
Cuba	Berkeley	44.7	22	3.32
Mexico	Berkeley	19.8	20	3.52
		23	20	3.65
		18	22	3.58
		22.5	24	3.75
		23.5	24	4.06
				3.71
		23	24	3.67
		24.3	24	2.57
		23.3	25	3.52
		15.2	25	3.66
		16	25	3.84
		25.1	25	3.77
		24.5	25	3.60
		23.3	25	3.53
		24.4	25	3.73
		24.7	25	3.81
		18	26	3.62
		24.4	26	3.87
		24	26	3.76
		20.1	27	3.95
		21	27	3.92
				3.81
		18.3	28	3.68
		19.4	28	3.99
		15.3	30	3.80
		17.3	30	3.90
		15.2	30	3.85
		19.8	30	3.67
		18.3	30	3.80
		15.3	30	3.85

TABLE 39.—(*Continued*)

Epicentral region	Station	Arcual distance, °	Observed wave period, sec.	Observed mean speed, km./sec.
Mexico (*Continued*)	Berkeley (*Continued*)	30	30	3.88
		29.5	30	3.57
		23.4	30	3.77
		32	30	3.77
		24.7	30	3.98
				3.91
		24	30	3.82
		23.1	30	3.81
		23.6	30	3.73
		29.2	33	3.72
		34.5	34	3.74
		15	35	4.02
		11.3	35	4.10
		26.3	35	3.90
		32	35	3.85
		30.7	36	3.92
		30.1	36	3.95
		31.2	36	3.92
		32.3	40	3.86
				3.93
		34	40	3.96
		33.4	40	4.04
		34.5	40	4.00
		34	40	3.88
		32	40	3.92
		22.6	40	3.67
		31.1	40	3.79
				3.87
				3.89
		24.4	40	3.77
		31.6	50	4.00
		31.1	60	4.12
	Mount Hamilton	30.5	45	3.90
		25.7	35	3.93
		30	34	3.99
		24.0	30	3.87
		31.8	30	3.82
		14.7	30	3.83
		10.7	30	3.81
		23.7	24	3.54
	Palo Alto	11.0	30	3.91
		30.2	35	3.99
				3.97
		31.6	40	3.94
		30.7	40	3.79
Panama	Berkeley	48.7	40	3.76
		44.4	40	3.97
		49.2	38	3.70
		48.7	35	3.63
		45	35	3.97
	Mount Hamilton	43.8	40	3.96
Puerto Rico	Berkeley	50.6	25	3.59
		50.5	28	3.47
		51	50	3.93
Venezuela	Berkeley	55.5	27	3.32
	Mount Hamilton	58.5	45	3.88

TABLE 40

VELOCITIES OF SURFACE SHEAR WAVES ACROSS THE AMERICAS

(After Gutenberg and Richter[71])

Epicentral region	Station	Arcual distance, °	Observed wave period, sec.	Observed mean speed, km./sec.
Alaska	Florissant	41.5	7	3.6
		42	30	3.5
			15	3.3
	Charlottesville	48	46	3.9
			36	3.7
	Chicago	40.5	10	3.6
	Halifax	49.5	9	3.6
	Harvard	47.5	55	4.3
			42	4.0
	Little Rock	44	10	3.5
	Ottawa	38.5	110	4.7
			60	4.4
		44	48	4.0
			36	3.7
			10	3.6
			40	4.0
			8	3.6
	St. Louis	41.5	6	3.8
		42	38	3.6
			30	3.5
	East Machias	48.5	48	4.1
Baffin Bay	Pasadena	46	70	4.0
			45	3.9
			50	4.0
			32	3.7
			24	3.5
	Riverside	46	8	3.7
	Santa Barbara	46.5	7	3.8
	Tinemaha	43	6	3.6
Chile	La Paz	17.5	12	3.5
	La Plata	12.5	30	3.6
		18	30	3.9
	Rio de Janeiro	25	10	3.6
		27.5	14	3.2
Colombia	La Paz	25	15	3.9
	La Plata	45.5	25	3.4
	Rio de Janeiro	42.5	8	3.6
Ecuador	Huancayo	11	20	3.2
	La Paz	18.5	13	3.5
	La Plata	38	22	3.4
	Rio de Janeiro	41	8	3.3
Mexico	Berkeley	23	18	3.3
	Berkeley and Lick	28.5	27	3.8
			22	3.8
	Florissant	23	7	3.3
		25	10	3.4
	Georgetown	31	30	3.7
	Harvard	36.5	12	3.3
	Chicago	28.5	16	3.4
			13	3.3

TABLE 40.—(*Continued*)

Epicentral region	Station	Arcual distance, °	Observed wave period, sec.	Observed mean speed, km./sec.
Mexico (*Continued*)	La Jolla	19	20	3.7
	Little Rock	19.5	4	3.4
	Mount Wilson	20	16	3.7
	Ottawa	34	40	3.9
			30	3.5
		35.5	8	3.5
		37.5	12	3.4
	Pasadena	7	50 ±	4.5
			35	4.1
			12	3.6
		17.5	20	3.7
		18.5	32	4.3
			22	4.0
		20	30	3.7
		11.5	3	4.2
		20	17	3.5
			24	3.6
			26	4.0
			25	4.0
		22	32	4.3
			25	4.1
		24	40	4.1
			22	3.8
		25.5	40	4.0
			35	3.9
			22	3.6
			40	4.0
			35	4.0
			23	3.8
	San Juan	36	44	4.1
			30	3.6
	Santa Barbara	19.5	17	3.5
	Saskatoon	33.5	24	3.5
	Sitka	44	44	4.0
			30	3.8
	Stanford	28.5	24	3.8
	Tinemaha	10	30	4.0
Nevada......................	Halifax	40.5	60	4.5
			30	3.8
	Ottawa	32	16	3.6
	San Juan	49.5	40	4.1
			32	3.8
	Saskatoon	15	6	3.2
	Sitka	21.5	40	4.1
			32	4.0
Newfoundland................	Haiwee	47	8	3.7
	Pasadena	48	6	3.6
	Victoria	45	60	4.4
			14	3.9
Nicaragua...................	Pasadena	37	42	3.9
			36	3.8
Panama......................	Pasadena	42	42	4.1
			38	4.0
			40	3.9

TABLE 40.—(*Continued*)

Epicentral region	Station	Arcual distance, °	Observed wave period, sec.	Observed mean speed, km./sec.
Peru	La Plata	24	42	3.7
			32	3.5
	Rio de Janeiro	38.5	60	4.1
Salvador	Pasadena	34	35	3.7
Texas	Chicago	17	19	3.5
			12	3.4
	Haiwee	12.5	20	3.6
			12	3.3
	Mount Wilson	12	17	3.5
	Ottawa	26.5	7	3.6
			8	3.5
	Sitka	34	23	3.5
			18	3.4
Utah	Georgetown	27.5	13	3.8
	Chicago	19	14	3.7
			12	3.6
	Little Rock	17.5	4	3.5
			6	3.7
			2	3.6
	Ottawa	27.5	16	3.9
	Sitka	21	28	3.4
			12	3.3

TABLE 41

VELOCITIES OF SURFACE SHEAR WAVES THROUGH EURASIA

(After Gutenberg[58])

Epicentral region	Station	Arcual distance, °	Observed wave period, sec.	Observed mean speed, km./sec.
Japan	Cartuja	99.8	24	3.27
			28	3.36
			40	3.59
	De Bilt	83.9	16	3.37
			17	3.00
			24	3.32
			30	3.43
			35	3.65
			40	3.63
			60–70	3.85
	Feldberg	84.1	17	3.09
			19	3.12
			24	3.12
				3.18
			25	3.32
			28	3.39
			30	3.51
			35	3.47
	Hamburg	80.9	16	3.26
			18	3.28
			20	3.34
			30	3.45
	Hohenheim	84.5	16	3.01
			17	3.16
			20	3.44
			60–70	4.22
	Jena	82.3	16	3.17
			20	3.09
	Munich	84.4	16	3.17
			17	3.29
			20	3.34
			24	3.41
			25	3.43
	Potsdam	80.6	19	3.26
			30	3.45
	Ravensburg	85.5	16	3.27
			60–70	4.26
	Upsala	73.5	16	2.98
			17	3.30
			22	3.47
			30	3.56
	Vienna	82.3	16	3.11
			18	3.27
			20	3.22
			22	3.25
			45	3.87
				3.90
			60–70	4.20
	Zurich	86.5	19	3.31
			20	3.44
			24	3.52

TABLE 42

VELOCITIES OF SURFACE SHEAR WAVES ACROSS THE PACIFIC BASIN EXCLUDING WEST AND SOUTHWEST

Epicentral region	Station	Arcual distance, °	Observed wave period, sec.	Observed mean speed, km./sec.	Authority
Alaska	Berkeley	26.6	20	4.13	Carder[35]
				3.93	Carder[35]
		26	26	3.95	Carder[35]
		30	30	4.20	Carder[35]
	Honolulu	39.2	60	4.66	Carder[35]
			45	4.53	Carder[35]
	Wellington	106.6	45	4.42	Carder[35]
Alaskan Peninsula	Berkeley	27.7	20	4.00	Carder[35]
		28.4	26	4.17	Carder[35]
			30	4.24	Carder[35]
		29.3	30	4.12	Carder[35]
		30.0	30	4.08	Carder[35]
Aleutian Islands	Berkeley	35.4	16	3.97	Carder[35]
			25	4.33	Carder[35]
		41.6	30	4.31	Carder[35]
		30.5	32	4.16	Carder[35]
		42.3	32	4.33	Carder[35]
		41.6	35	4.36	Carder[35]
		39.5	35	4.34	Carder[35]
		42.5	35	4.34	Carder[35]
		40.3	35	4.39	Carder[35]
			36	4.41	Carder[35]
		35.5	40	4.53	Carder[35]
	La Jolla	53	31	4.3	Gutenberg and Richter[71]
	Mount Hamilton	36.1	36	4.40	Carder[35]
		47.1	34	4.32	Carder[35]
		40.1	25	4.15	Carder[35]
	Palo Alto	46.5	35	4.32	Carder[35]
		37.0	36	4.37	Carder[35]
	Pasadena	51	27	4.2	Gutenberg and Richter[71]
			23	4.3	Gutenberg and Richter[71]
		44	17	4.1	Gutenberg and Richter[71]
			16	4.1	Gutenberg and Richter[71]
		43	21	4.3	Gutenberg and Richter[71]
			12	4.1	Gutenberg and Richter[71]
		40.5	36	4.4	Gutenberg and Richter[71]
			32	4.3	Gutenberg and Richter[71]
		39.5	19	4.2	Gutenberg and Richter[71]
			18	4.4	Gutenberg and Richter[71]
Bismarck Archipelago	Huancayo	129.5	48	4.4	Gutenberg and Richter[71]
		130	60	4.4	Gutenberg and Richter[71]
			70	4.4	Gutenberg and Richter[71]
		134	60	4.3	Gutenberg and Richter[71]
			32	4.5	Gutenberg and Richter[71]
	Pasadena	93	35	4.3	Gutenberg and Richter[71]
		92	35	4.4	Gutenberg and Richter[71]
		94	40	4.5	Gutenberg and Richter[71]
California	Apia	69	30	4.07	Carder[35]
	Hawaii	34.4	30	4.16	Carder[35]
	Wellington	96.8	30	4.09	Carder[35]

TABLE 42.—(*Continued*)

Epicentral region	Station	Arcual distance, °	Observed wave period, sec.	Observed mean speed, km./sec.	Authority
Caroline Islands	Berkeley	89.6	24	4.00	Carder[35]
	Mount Hamilton	90	30	4.10	Carder[35]
Chile	Apia	93	90	4.66	Carder[35]
	Honolulu	98.6	100	4.76	Carder[35]
	Tokyo	154.5	160	4.80	Carder[35]
	Wellington	83	30	4.4	Gutenberg and Richter[71]
Easter Island	Huancayo	38	16	4.4	Gutenberg and Richter[71]
	Pasadena	58	21	4.4	Gutenberg and Richter[71]
			13	4.2	Gutenberg and Richter[71]
		61	21	4.3	Gutenberg and Richter[71]
			8	4.3	Gutenberg and Richter[71]
	Victoria	75	40	4.3	Gutenberg and Richter[71]
Fiji Islands	Berkeley	76.5	30	4.14	Carder[35]
		79.3	32	4.10	Carder[35]
		76.5	34	4.23	Carder[35]
			35	4.18	Carder[35]
		78.6	50	4.50	Carder[35]
	Mount Hamilton	78.6	30	4.21	Carder[35]
Galápagos Islands	Huancayo	31.5	10	4.3	Gutenberg and Richter[71]
Guam	Berkeley	85.7	35	4.51	Carder[35]
Hawaii	Berkeley	34.6	26	4.16	Carder[35]
	Pasadena	36	15	4.0	Gutenberg and Richter[71]
	Victoria	38	15	4.2	Gutenberg and Richter[71]
Japan	Berkeley	70.4	60	4.62	Carder[35]
		75	60	4.66	Carder[35]
		77.3	43	4.50	Carder[35]
		82	40	4.51	Carder[35]
		75+	38	4.02	Carder[35]
		75	38	4.49	Carder[35]
		82	36	4.44	Carder[35]
		65	35	4.45	Carder[35]
		75.2	30	4.35	Carder[35]
		72.7	30	4.43	Carder[35]
		82	30	4.26	Carder[35]
		70	27	4.00	Carder[35]
			26	4.00	Carder[35]
		74	25	4.39	Carder[35]
		74.8	16	4.08	Gutenberg[58]
			18	4.22	Gutenberg[58]
			20	4.22	Gutenberg[58]
			22	4.27	Gutenberg[58]
			32	4.34	Gutenberg[58]
	Honolulu	55.6	40	4.47	Carder[35]
		55.9	17	4.06	Gutenberg[58]
			24	4.32	Gutenberg[58]
	La Paz	149.0	20	3.85	Gutenberg[58]
			35	3.86	Gutenberg[58]
			40	4.34	Gutenberg[58]
	Mount Hamilton	75.6	16	4.00	Gutenberg[58]
			17	4.17	Gutenberg[58]
			22	4.30	Gutenberg[58]
			30	4.37	Gutenberg[58]
		75.5	60	4.65	Carder[35]
		78	50	4.56	Carder[35]

TABLE 42.—(*Continued*)

Epicentral region	Station	Arcual distance, °	Observed wave period, sec.	Observed mean speed, km./sec.	Authority
Japan (*Continued*).....	Pasadena	75.4	190	4.78	Carder[35]
		75	23	4.4	Gutenberg and Richter[71]
			14	4.3	Gutenberg and Richter[71]
			90	4.6	Gutenberg and Richter[71]
			18	4.4	Gutenberg and Richter[71]
			32	4.4	Gutenberg and Richter[71]
			14	4.2	Gutenberg and Richter[71]
		81	26	4.4	Gutenberg and Richter[71]
		87	25	4.2	Gutenberg and Richter[71]
	Santa Barbara	74	18	4.4	Gutenberg and Richter[71]
	Tinemaha	72	23	4.4	Gutenberg and Richter[71]
			40	4.5	Gutenberg and Richter[71]
Japan Sea.............	Berkeley	88	50	4.62	Carder[35]
Kamchatka...........	Berkeley	52.7	26	3.89	Carder[35]
		52.5	27	4.16	Carder[35]
		50.5	30	3.90	Carder[35]
		56.4	30	4.39	Carder[35]
		55.2	32	4.01	Carder[35]
		52.7	24.8	3.93	Carder[35]
		52	35	4.26	Carder[35]
		53+	80	4.66	Carder[35]
	Mount Hamilton	57	54	4.49	Carder[35]
			50	4.58	Carder[35]
			30	4.36	Carder[35]
	Pasadena	63	20	4.3	Gutenberg and Richter[71]
Kermadec Islands......	Berkeley	84.5	60	4.51	Carder[35]
		86+	30	4.08	Carder[35]
		86	30	4.03	Carder[35]
		84.5	30	4.07	Carder[35]
			60	4.21	Carder[35]
	Mount Hamilton	84.2	27	4.39	Carder[35]
Kuril Islands..........	Berkeley	65	30	4.37	Carder[35]
		63	30	4.43	Carder[35]
		59+	30	4.41	Carder[35]
		61.8	44	4.45	Carder[35]
		60.6	36	4.42	Carder[35]
		60−	35	4.31	Carder[35]
	Mount Hamilton	61.8	60	4.68	Carder[35]
		62.4	48	4.62	Carder[35]
		60−	30	4.38	Carder[35]
Ladrones..............	Berkeley	86	80	4.68	Carder[35]
			50	4.57	Carder[35]
		79.4	40	4.41	Carder[35]
		79.5	35	4.12	Carder[35]
		79	30	4.11	Carder[35]
		79.5	27	4.04	Carder[35]
	Pasadena	84	21	4.4	Gutenberg and Richter[71]
		84.5	20	4.5	Gutenberg and Richter[71]
Mexico..............	Honolulu	46.5	15	4.3	Gutenberg and Richter[71]
			25	4.5	Gutenberg and Richter[71]
		50	8	4.0	Gutenberg and Richter[71]
			15	4.4	Gutenberg and Richter[71]

TABLE 42.—(*Continued*)

Epicentral region	Station	Arcual distance, °	Observed wave period, sec.	Observed mean speed, km./sec	Authority
New Hebrides..........	Berkeley	86.7	20	3.97	Carder[35]
		85	30	4.11	Carder[35]
		84.5	30	4.07	Carder[35]
		86	30	4.09	Carder[35]
				4.08	Carder[35]
		84.7	30	4.13	Carder[35]
		80	36	4.04	Carder[35]
		86.5	36	4.17	Carder[35]
		86+	38	4.17	Carder[35]
		85	40	4.14	Carder[35]
		87.7	40	4.12	Carder[35]
		82.7	40	4.21	Carder[35]
		86	50	4.43	Carder[35]
	Huancayo	112	24	4.3	Gutenberg and Richter[71]
Samoa................	Berkeley	74	30	4.27	Carder[35]
		72.5	30	4.03	Carder[35]
	Huancayo	93	40	4.5	Gutenberg and Richter[71]
Santa Cruz Islands.....	Huancayo	114	65	4.4	Gutenberg and Richter[71]
	Pasadena	84.5	40	4.5	Gutenberg and Richter[71]
		85	22	4.3	Gutenberg and Richter[71]
Solomon Islands........	Berkeley	90.5	40	4.08	Carder[35]
		89.5	50	4.52	Carder[35]
		92	48	4.31	Carder[35]
		87	48	4.50	Carder[35]
		88.7	48	4.50	Carder[35]
		87.7	48	4.49	Carder[35]
				4.46	Carder[35]
		92	40	4.26	Carder[35]
		88.5	40	4.15	Carder[35]
		86.5	40	4.18	Carder[35]
		88.8	40	4.22	Carder[35]
		88	40	4.12	Carder[35]
		89	40	4.20	Carder[35]
		88.5	37	4.15	Carder[35]
		89.5	36	4.36	Carder[35]
		86	36	4.17	Carder[35]
		87.5	35	4.20	Carder[35]
		86.5	30	4.06	Carder[35]
		87.5	29	4.14	Carder[35]
		90+	26	4.07	Carder[35]
		85.5	45	4.5	Gutenberg and Richter[71]
			19	4.4	Gutenberg and Richter[71]
	Haiwee	88.5	52	4.5	Gutenberg and Richter[71]
	Hawaii	52	36	4.5	Gutenberg and Richter[71]
			33	4.2	Gutenberg and Richter[71]
	Honolulu	51	11	4.0	Gutenberg and Richter[71]
			24	4.5	Gutenberg and Richter[71]
	Huancayo	116	40	4.3	Gutenberg and Richter[71]
			28	4.3	Gutenberg and Richter[71]
		119	48	4.4	Gutenberg and Richter[71]
	La Jolla	88.5	48	4.5	Gutenberg and Richter[71]

TABLE 42.—(*Continued*)

Epicentral region	Station	Arcual distance, °	Observed wave period, sec.	Observed mean speed, km./sec.	Authority
Solomon Islands (*Continued*)	La Paz	124	100	4.6	Gutenberg and Richter[71]
			60	4.5	Gutenberg and Richter[71]
		133.5	60	4.42	Carder[35]
			50	4.32	Carder[35]
	Mount Hamilton	92+	48	4.31	Carder[35]
		86	45	4.5	Gutenberg and Richter[71]
			20	4.4	Gutenberg and Richter[71]
	Mount Wilson	88	50	4.5	Gutenberg and Richter[71]
			12	4.2	Gutenberg and Richter[71]
	Pasadena	88	25	4.5	Gutenberg and Richter[71]
			22	4.4	Gutenberg and Richter[71]
		91	35	4.6	Gutenberg and Richter[71]
			10	4.0	Gutenberg and Richter[71]
		86.5	25	4.5	Gutenberg and Richter[71]
			14	4.3	Gutenberg and Richter[71]
		88	18	4.4	Gutenberg and Richter[71]
			46	4.5	Gutenberg and Richter[71]
			17	4.4	Gutenberg and Richter[71]
	Riverside	88.5	50	4.5	Gutenberg and Richter[71]
	Santa Barbara	86.5	54	4.5	Gutenberg and Richter[71]
	Santa Clara	86	38	4.1	Gutenberg and Richter[71]
	Santiago	114	70	4.5	Gutenberg and Richter[71]
	Seattle	88.5	52	4.5	Gutenberg and Richter[71]
	Sitka	85	60	4.5	Gutenberg and Richter[71]
			30	4.5	Gutenberg and Richter[71]
			16	4.0	Gutenberg and Richter[71]
				4.4	Gutenberg and Richter[71]
	Stanford	85.5	30	4.4	Gutenberg and Richter[71]
			18	4.3	Gutenberg and Richter[71]
	Tinemaha	88.5	48	4.5	Gutenberg and Richter[71]
	Ukiah	85.5	28	4.5	Gutenberg and Richter[71]
	Victoria	88	53	4.5	Gutenberg and Richter[71]
			45	4.5	Gutenberg and Richter[71]

TABLE 43

VELOCITIES OF SURFACE SHEAR WAVES ACROSS THE WESTERN AND SOUTHWESTERN PACIFIC OCEAN

Epicentral region	Station	Arcual distance, °	Observed wave period, sec.	Observed mean speed, km./sec.	Authority
Japan......	Batavia	51.6	22	3.39	Gutenberg[58]
			30	3.45	Gutenberg[58]
			35	3.91	Gutenberg[58]
	Malabar	50.8	22	3.04	Gutenberg[58]
			60–70	4.21	Gutenberg[58]
	Manila	26.5	16	3.47	Gutenberg[58]
					Gutenberg and Richter[71]
Solomon Islands...	Hong Kong	57	37	4.2	Gutenberg and Richter[71]
	Kobe	51.5	54	4.3	Gutenberg and Richter[71]
			47	4.3	Gutenberg and Richter[71]
	Manila	47.5	19	3.5	Gutenberg and Richter[71]
	Phu Lien	62.5	41	4.2	Gutenberg and Richter[71]
			25	3.7	Gutenberg and Richter[71]
	Tokyo	50.5	54	4.1	Gutenberg and Richter[71]
			45	4.0	Gutenberg and Richter[71]
			23	3.8	Gutenberg and Richter[71]
	Zikawei	56.5	55	4.4	Gutenberg and Richter[71]
			22	3.9	Gutenberg and Richter[71]

TABLE 44

VELOCITIES OF SURFACE SHEAR WAVES ACROSS THE NORTH POLAR REGIONS

Epicentral region	Station	Arcual distance, °	Observed wave period, sec.	Observed mean speed, km./sec.	Authority
Alaska......	Abisko	51	36	4.4	Gutenberg and Richter[71]
		53	36	4.1	Gutenberg and Richter[71]
		56	26	4.0	Gutenberg and Richter[71]
	Copenhagen	63.9	60	4.45	Carder[35]
			55	4.37	Gutenberg and Richter[71]
	Helsingfors	57.5	40	4.3	Gutenberg and Richter[71]
		64.5	24	4.3	Gutenberg and Richter[71]
	Kew	65.3	70	4.19	Carder[35]
	Leningrad	61.9	60	4.46	Carder[35]
	Oxford	65	60	4.28	Carder[35]
	Pulkovo	62	65	4.42	Carder[35]
	Upsala	58	45	4.2	Gutenberg and Richter[71]
		62	36	4.2	Gutenberg and Richter[71]
		64.5	40	4.1	Gutenberg and Richter[71]
		60.7	60	4.38	Carder[35]

TABLE 45

VELOCITIES OF SURFACE SHEAR WAVES ACROSS THE ATLANTIC OCEAN (After Gutenberg and Richter[71])

Epicentral region	Station	Arcual distance, °	Observed wave period, sec.	Observed mean speed, km./sec.
Central Atlantic..........	Harvard	69.5	28	4.1
	San Juan	56	32	4.1
			20	4.0
	Scoresby Sound	60	32	4.3
			19	4.2
	Technology	66.5	22	4.4
East Atlantic............	San Juan	47	40	4.5
			32	4.3
	Scoresby Sound	31.5	22	4.0
North Atlantic...........	Ivigtut	8.5	12	3.8
	Scoresby Sound	15.5	11	3.9
			9	3.7
South Atlantic...........	La Plata	33	25	4.2
			40	4.5
		38	22	4.2
		39.5	22	4.4
	Rio de Janeiro	38.25	28	4.1
		41.5	20	4.0
		43	18	4.1
	Scoresby Sound	124	30	4.3

TABLE 46

VELOCITIES OF SURFACE SHEAR WAVES ACROSS THE INDIAN OCEAN (After Gutenberg and Richter[71])

Epicentral region	Station	Arcual distance, °	Observed wave period, sec.	Observed mean speed, km./sec.
Madagascar................	Adelaide	65	33	4.0
	Bombay	53.5	20	4.1
				4.0
	Colombo	46	16	3.9
	Melbourne	69.5	45	4.4
	Perth	49	11	4.2
			14	4.0
	Wellington	86	50	4.4
			30	4.3
Solomon Islands............	Johannesburg	74	25	4.0
			33	4.2
			45	4.1
	Tananarive	55.5	29	4.5
			25	4.4

TABLE 47

Velocity of Pseudo-Rayleigh Waves across the Americas

(After Gutenberg and Richter[71])

Epicentral region	Station	Arcual distance, °	Observed wave period, sec.	Observed mean speed, km./sec.
Alaska	St. Louis	42	17	2.9
Baffin Bay	Pasadena	46	70	4.0
			45	3.9
			22	3.5
Chile	La Plata	12.5	30	3.2
		18	18	2.8
Colombia	La Paz	25	15	3.0
	Rio de Janeiro	42.5	10	3.1
Ecuador	Huancayo	11	12	2.8
	La Plata	38	30	2.7
Honduras	Pasadena	34	40	3.8
Mexico	Florissant	25	30	3.4
	Georgetown	31	24	3.3
	Harvard	36.5	8	2.9
	La Jolla	19	17	3.4
	Ottawa	34	45	3.8
	Pasadena	17.5	25	3.6
			16	3.3
		18.5	30	3.7
			22	3.6
	St. Louis	25	12	3.1
Nicaragua	Pasadena	37	30	3.4
			25	3.4
Panama	Pasadena	42	25	3.4
			30	3.4
Peru	La Plata	24	15	3.1
Texas	Charlottesville	22	24	3.4
Utah	Florissant	18	12	3.4
			7	3.1
	Georgetown	27.5	8	3.2
	St. Louis	18	10	3.1

TABLE 48

VELOCITIES OF PSEUDO-RAYLEIGH WAVES ACROSS EURASIA

(After Gutenberg[58])

Epicentral region	Station	Arcual distance, °	Observed wave period, sec.	Observed mean speed, km./sec.
Japan...............	Cartuja	99.8	16	2.91
			20	3.11
	De Bilt	83.9	13	2.79
			15	2.86
			24	3.15
			16	2.96
			20	3.02
	Feldberg	84.1	13	2.74
				2.81
	Hamburg	80.9	13	2.76
			15	2.70
				2.94
			16	2.91
	Hohenheim	84.5	17	2.66
			36	3.85
	Jena	82.3	13	2.71
				2.74
			14	2.81
			16	2.92
	Munich	84.4	13	2.81
			14	2.85
	Potsdam	80.6	13	2.68
	Ravensburg	85.5	14	2.69
	Upsala	73.5	14	2.66
			17	2.83
	Zurich	86.5	14	2.83
			18	2.96

TABLE 49

VELOCITIES OF PSEUDO-RAYLEIGH WAVES ACROSS THE PACIFIC EXCLUDING WEST AND SOUTHWEST

Epicentral region	Station	Arcual distance, °	Observed wave period, sec.	Observed mean speed, km./sec.	Authority
Aleutian Islands	La Jolla	53	25	4.0	Gutenberg and Richter[71]
	Pasadena	44	28	4.0	Gutenberg and Richter[71]
		40.5	25	4.0	Gutenberg and Richter[71]
			17	3.9	Gutenberg and Richter[71]
Bismarck Islands	Pasadena	92	29	4.0	Gutenberg and Richter[71]
			34	4.1	Gutenberg and Richter[71]
			22	3.9	Gutenberg and Richter[71]
		93	35	4.0	Gutenberg and Richter[71]
			26	3.9	Gutenberg and Richter[71]
		94	30	4.0	Gutenberg and Richter[71]
		95	35	4.0	Gutenberg and Richter[71]
Bonin Islands	Huancayo	140	32	4.0	Gutenberg and Richter[71]
Chile	Wellington	83	30	4.0	Gutenberg and Richter[71]
		85.5	35	4.1	Gutenberg and Richter[71]
		93.5	30	4.0	Gutenberg and Richter[71]
			24	4.0	Gutenberg and Richter[71]
Easter Island	Berkeley	64	27	3.9	Gutenberg and Richter[71]
	Huancayo	38	12	3.9	Gutenberg and Richter[71]
Galápagos Islands	Pasadena	41.5	22	3.9	Gutenberg and Richter[71]
Hawaii	Pasadena	36	21	4.0	Gutenberg and Richter[71]
Japan	Berkeley	74.8	16	2.68(?)	Gutenberg[58]
	Huancayo	137	33	3.9	Gutenberg and Richter[71]
	Honolulu	55.9	15	2.38	Gutenberg[58]
			17	3.09	Gutenberg[58]
			24	3.91	Gutenberg[58]
			27	4.32	Gutenberg[58]
	La Paz	149.0	20	3.43	Gutenberg[58]
			21	3.52	Gutenberg[58]
			27	3.71	Gutenberg[58]
	Pasadena	75	25	4.0	Gutenberg and Richter[71]
		89	36	4.1	Gutenberg and Richter[71]
			32	4.1	Gutenberg and Richter[71]
Kamchatka	Pasadena	57	30	4.0	Gutenberg and Richter[71]
			23	3.7	Gutenberg and Richter[71]
		67	20	3.8	Gutenberg and Richter[71]
Kermadec Islands	Pasadena	84.5	33	4.0	Gutenberg and Richter[71]
			20	3.7	Gutenberg and Richter[71]
Mexico	Honolulu	50	25	4.0	Gutenberg and Richter[71]
		58	30	4.1	Gutenberg and Richter[71]
New Hebrides	Huancayo	112	30	3.8	Gutenberg and Richter[71]
				4.0	Gutenberg and Richter[71]
	Pasadena	83	32	4.0	Gutenberg and Richter[71]
		86	24	4.0	Gutenberg and Richter[71]
		87	25	4.0	Gutenberg and Richter[71]
Samoa	Berkeley.	69	30	4.1	Gutenberg and Richter[71]
	Huancayo	96	24	4.0	Gutenberg and Richter[71]
	Pasadena	71	30	4.0	Gutenberg and Richter[71]
			34	4.0	Gutenberg and Richter[71]
			24	3.9	Gutenberg and Richter[71]
			21	3.9	Gutenberg and Richter[71]
		72	25	4.1	Gutenberg and Richter[71]
Santa Cruz Islands	Pasadena	84.5	35	4.1	Gutenberg and Richter[71]
				4.2	Gutenberg and Richter[71]
		85	36	4.0	Gutenberg and Richter[71]
Solomon Islands	Berkeley	85.5	34	4.1	Gutenberg and Richter[71]
	Huancayo	116	30	3.9	Gutenberg and Richter[71]
		119	30	4.0	Gutenberg and Richter[71]
	La Jolla	88.5	30	4.0	Gutenberg and Richter[71]
	Pasadena	88	35	4.0	Gutenberg and Richter[71]
			28	4.0	Gutenberg and Richter[71]
			25	3.9	Gutenberg and Richter[71]
			40	4.1	Gutenberg and Richter[71]
			30	4.0	Gutenberg and Richter[71]
			33	4.1	Gutenberg and Richter[71]
			27	4.0	Gutenberg and Richter[71]
		91	33	4.2	Gutenberg and Richter[71]
			22	4.1	Gutenberg and Richter[71]
			25	3.9	Gutenberg and Richter[71]
		91.5	35	4.2	Gutenberg and Richter[71]
				4.0	Gutenberg and Richter[71]
		86.5	30	4.1	Gutenberg and Richter[71]
			24	4.0	Gutenberg and Richter[71]
	Santa Barbara	86.5	33	4.1	Gutenberg and Richter[71]
			32	4.1	Gutenberg and Richter[71]
	Tinemaha	88.5	35	4.0	Gutenberg and Richter[71]

TABLE 50

VELOCITIES OF PSEUDO-RAYLEIGH WAVES ACROSS THE WESTERN AND SOUTHWESTERN PACIFIC OCEAN

Epicentral region	Station	Arcual distance, °	Observed wave period, sec.	Observed mean speed, km./sec.	Authority
Japan..........	Batavia	51.6	14	2.45	Gutenberg[58]
			23	3.11	Gutenberg[58]
Solomon Islands.	Apia	36	22	3.7	Gutenberg and Richter[71]
			7	3.6	Gutenberg and Richter[71]
	Kobe	51.5	19	3.5	Gutenberg and Richter[71]
	Riverview	25	33	3.9	Gutenberg and Richter[71]
	Tokyo	50.5	50	3.8	Gutenberg and Richter[71]
			18	3.5	Gutenberg and Richter[71]

TABLE 51

VELOCITIES OF PSEUDO-RAYLEIGH WAVES ACROSS THE NORTH POLAR REGIONS (After Gutenberg and Richter[71])

Epicentral region	Station	Arcual distance, °	Observed wave period, sec.	Observed mean speed, km./sec.
Alaska..................	Abisko	51	30	3.7
		56	24	3.7
	Scoresby Sound	45	30	3.5
	Upsala	58	30	3.8
		64.5	28	3.6

TABLE 52

VELOCITIES OF PSEUDO-RAYLEIGH WAVES ACROSS THE ATLANTIC OCEAN (After Gutenberg and Richter[71])

Epicentral region	Station	Arcual distance, °	Observed wave period, sec.	Observed mean speed, km./sec.
North Atlantic............	Harvard	27	22	3.4
	Ivigtut	8.5	20	3.7
			12	3.4
	Scoresby Sound	15.5	8	3.3
South Atlantic............	La Plata	33	30	3.9
	Scoresby Sound	124	22	3.7

TABLE 53
VELOCITIES OF PSEUDO-RAYLEIGH WAVES ACROSS THE INDIAN OCEAN
(After Gutenberg and Richter[71])

Epicentral region	Station	Arcual distance, °	Observed wave period, sec.	Observed mean speed, km./sec.
Madagascar................	Batavia	53	23	3.9
			20	4.1
	Bombay	53.5	22	3.9
				3.8
	Christchurch	83	30	4.0
			27	3.9
	Melbourne	69.5	24	3.9
	Tananarive	55.5	27	4.0
	Wellington	86	27	3.9
			30	4.0

SEISMIC EVIDENCE FOR THE CONSTITUTION OF THE DEEPER SHELLS BETWEEN THE DEEPEST CRUSTAL DISCONTINUITY AND THE OUTER BOUNDARY OF THE CORE

With the advent of precise seismometry came the discovery that the material which constitutes the body of the earth underneath the crust is able to transmit shear waves with such speed as to indicate a high degree of rigidity. In order to account for the apparent disappearance and the marked retardation on reappearance near the antipodes of the condensation waves as they are followed around the earth, Oldham[162] and Wiechert[210] adopted the hypothesis of a mantle and a core.

At the outer boundary of the mantle beneath the Mohorovičić discontinuity the velocities of condensation and shear waves have been tabulated above as P_n and S_n, respectively. However, as these velocities involve the square root of the ratio of the elasticity to the density, as is stated at the beginning of this chapter, it would be impossible to distinguish between materials of quite different density provided that the modulus of incompressibility and the modulus of rigidity varied with the density. Hence, as Daly[42,43,44,45,46] has pointed out, we should probably not be able at present to distinguish between holocrystalline gabbro and glassy basalt by means of the observed travel times of seismic waves unless the decrease in velocity were sufficient to cause the phenomenon of a shadow zone.

The determination of the velocities of the condensation waves P and of the shear waves S and, consequently, of the structure at depths much greater than the Mohorovičić discontinuity is a problem that

depends for its solution on the construction of exact time-distance graphs or travel-time curves for earthquakes of all depths of focus. These depend on three distinct sets of empirical data: (1) the determination of the initial conditions, especially the depth of focus; (2) the precise observation of the arrival times at successive points along the curved surface of the earth; (3) the working out of the crustal structure at the epicenter and at each of the observing stations. Most of the older travel-time curves, such as those of Milne, of Wiechert and Zoeppritz and of Mohorovičić involved systematic errors. They were averages that ignored the initial and final conditions and supposed the earthquakes to have occurred in the surface of the earth. Even the recent tables of Jeffreys[94,96] and of Gutenberg and Richter[71] are statistical averages and are affected by unknown boundary conditions. Much information has been gleaned from such statistical averages. Also, it is possible by judicious smoothing to bring about a satisfactory agreement[73] between these averages and individual earthquakes. But if the differences are real, then very valuable information is being lost by the smoothing. There seems to be evidence[121,175,177] that the travel times of earthquakes in some regions differ from those in others by amounts which exceed reasonable limits of observational error and that these differences between individual earthquakes are greater than the differences between modern tables such as those of Jeffreys,[101] Jeffreys[104] and Bullen, Gutenberg[71] and Richter, and the present writer.[134,140] Therefore it would seem evident that further progress in our knowledge of structure will depend (1) on our ability to measure with increasing precision at a larger number of stations distributed throughout the whole range of epicentral distances the arrival times of the phases in one individual earthquake and (2) on our success in determining, with reliability and independently of any assumptions or of previously existing travel-time curves, the geographic location and depth of the focus of each individual earthquake chosen for investigation, the crustal structure of its focal region and of the areas in which the observations were made and the precise time at which the energy carried by the particular earthquake waves in question was radiated from the focus. These requirements presuppose a large number of high-sensitivity open-scale seismographs and a high-precision time service in the epicentral region.

Individual travel-time curves were constructed by Oldham for the Assam earthquake of June 12, 1897, by Omori for the Kangra earthquake of Apr. 4, 1905, by Reid for the California earthquake of Apr. 18, 1906, by Rizzo[169] for the Messina earthquake of Dec. 28, 1908, and by others. But neither the high instrumental sensitivity nor the

precise absolute timing of phases now known to be required was then available, nor was it possible in those days to determine the focal depth and crustal structure. Nevertheless, Rizzo's curve for the Messina earthquake was surprisingly good throughout a considerable part of its range.

The Tango (Japan) earthquake of Mar. 7, 1927, occurring in the midst of the closely spaced Japanese network of seismographic stations and recorded by many modern seismographs at considerable distances, offered Hodgson[83,84,85,87] an opportunity to construct travel-time curves based on definitely known initial conditions. The P curve was fixed with considerable precision in the distance intervals 2 to 8°, 50 to 100°. Dahm[39] showed that the focal depth of the extremely violent Hawke Bay (New Zealand) earthquake of Feb. 2, 1931, was of the same order as that of the Tango earthquake and that the two travel-time curves coincided within the range of overlap. Hence he concluded that the rest of the Hawke Bay curve, from 100 to 180°, could be taken as an extension of the Tango curve. Bullen's[25] study of New Zealand earthquakes, as stated above, tends to confirm this conclusion since it suggests a crustal structure similar to that of the Tango region. Since the depth of focus of the Long Beach (California) earthquake of Mar. 10, 1933, had been found by Wood and by Gutenberg to be 10 km. and, therefore, of the same order as that of the Tango earthquake, Dahm and the present writer[134] studied the original records at epicentral distances between 8 and 50° and constructed a travel-time curve, basing it on the epicenter and focal time determined by Wood[215] from the records of the southern California stations. This curve was found to connect at both ends of the above interval with the Tango observations although the crustal structure in California is different. The three curves together form an almost unbroken continuum covering the whole range of epicentral distance quite independently of all previous curves.

EVIDENCE FROM DEEP EARTHQUAKES

Studies by Wadati,[205] Scrase,[180,181,182] Stoneley,[193] Stechschulte,[191,192] Brunner,[22,23,24] and Gutenberg[70,72] and Richter have shown conclusively that earthquakes occur at depths ranging up to one-tenth of the earth's radius and that, as was expected, travel times are greatly influenced by focal depth.

The term *deep-focus earthquake* is an awkward one, but seismologists have not yet agreed upon a suitable substitute. In Bulletin 90 of this series the present writer coined and has since used consistently the term *plutonic earthquake* because the word plutonic in itself means *deep in the*

earth; it originally signified *pertaining to Pluto or to his kingdom of Hades* (which was supposed by the ancients to be in the interior of the earth), hence *belonging to the netherworld, deeply subterranean.* Some have objected to the word plutonic because it could be misunderstood by geologists as denoting or at least suggesting a connection between deep-focus earthquakes and the *old Plutonic theory*, or the doctrine of the *old Plutonists* that the phenomena of rock structure are chiefly due to deep-seated igneous activity, as opposed to the *old Neptunian theory. Modern* geologists retain the term *plutonic rock* but oppose it to *volcanic rock*, both types being of igneous origin but the one formed at great depth, the other at or near the surface. Although there is no intrinsic reason why the terminology of one science should be bound by that of another, nevertheless there is an obvious parallelism between the *modern* geological usage which gives the name plutonic to igneous rocks of *deep* origin as distinguished from those of *shallow* origin and the seismological application of the name plutonic to earthquakes of *deep* origin in contrast with *tectonic* and *volcanic* earthquakes which have a comparatively shallow origin. The present writer defines a *plutonic earthquake* as *one that has its focus at any depth whatever below the Mohorovičić discontinuity* at the base of the crustal layers. By the Mohorovičić discontinuity we mean that discontinuity which is responsible for the refraction of the P_n and S_n phases in the near-station records of those earthquakes which originate *above* the Mohorovičić discontinuity. This latter class includes volcanic and tectonic earthquakes. Volcanic earthquakes are usually very shallow. Tectonic earthquakes are said to be *normal* if their focus lies 10 to 15 km. below the surface of the earth. By the term *focus* we mean the origin or source of the *first* P and S waves of a given earthquake that arrive at the observing stations.

No method is known for the direct observation of focal depth as such. What is observed is the time of arrival of successive phases at stations scattered over the surface of the earth. From these arrival times, which are subject to the severe limitations we have mentioned above, we calculate by one of several methods the time required for a wave of P- or S-type to travel vertically upward from the focus to the epicenter. The problem of translating this *time* into a *depth* is the same as in reflection prospecting. By some means or other the actual *average velocity* between the focus and the surface must be found. It may be calculated from the time-distance graphs and the crustal structure provided that these can be found with sufficient precision. All the depths we have at present are approximations only. Nevertheless they represent a great advance over what was known in preceding

decades and doubtless give us the correct order of depth at least. Plutonic earthquakes constitute a powerful tool not only for the investigation of crustal structure but also for that of the deep interior. Two facts stand out. Plutonic earthquakes may be very strong and their energy may be released in the form of shear-wave energy. Therefore we may legitimately conclude that the material down to a depth of at least 700 km. is capable of storing great amounts of elastic-strain energy and that this potential energy may be stored largely in the form of *shearing strain* (for details see Chap. XI).

VELOCITY AS A FUNCTION OF DEPTH

Two ingeniously effective, though laborious, direct methods have been devised for the determination of the velocity of seismic body waves as a function of depth in the earth when the *true surface velocity* and the *apparent surface velocity* are known. The one is due to Knott[109] and depends on the solution of Abel's integral equation by Bateman.[13] The other is that of Wiechert and Geiger[213] and depends on the solution of the same equation by Herglotz.[79] The Bateman-Knott equation is

$$f(p) = p\int_p^{\frac{1}{V}} \frac{\partial/\partial u\,(\log r)\,du}{(u^2 - p}$$

and its solution is

$$\frac{\partial}{\partial u}(\log r) = -\frac{2}{\pi}\frac{\partial}{\partial u}\int_u^{\frac{1}{V}} \frac{f(p)\,dp}{(p^2 - u^2)^{\frac{1}{2}}}, \tag{39}$$

or, since the constant vanishes,

$$\log r = -\frac{2}{\pi}\int_u^{\frac{1}{V}} \frac{f(p)\,dp}{(p^2 - u^2)^{\frac{1}{2}}}, \tag{40}$$

in which p is a ray parameter equal to the apparent surface velocity or the rate at which the travel time changes with the arcual distance from the epicenter measured along the curved surface of the earth; $p = d\Delta/dT$, a constant for any given ray; v is the true velocity of the seismic wave in radians per second at any point in the earth whose distance r from the center is expressed as a fraction of the total radius R of the earth; V is the value of v at the surface of the earth; $u = r/v$.

The Herglotz-Wiechert equation is

$$\frac{\Delta}{2R} = \frac{1}{V_\Delta}\int_{V_l}^{v}$$

and its solution as transformed by Wiechert is

in which R is the total radius of the earth, r the radius vector from the center of the earth, V_Δ the apparent surface velocity at the arcual epicentral distance Δ, V_r the apparent surface velocity at the point of emergence of a ray whose deepest point or vertex is at the distance r from the center of the earth; Δ_r is the arcual distance to the point of emergence of this ray and $q = \cosh^{-1} (V_r/V_\Delta)$.

Herglotz had stated the basic requirement that the time-distance curve must be everywhere concave to the distance axis if the problem is to be reduced to Abel's equation. Witte[214] stated the further conditions: (1) that the earth may be considered a sphere; (2) that the origin of the earthquake may be considered to be a point in the surface; (3) that the elastic wave propagation can be treated by the two-dimensional ray method of optics; (4) that Fermat's principle is valid; (5) that each ray has one and only one point in which the tangent is perpendicular to the radius vector drawn from the center of the earth to the point and that the ray is symmetrical with regard to that point; (6) that the velocity of wave propagation is a function of the central radius vector only and is not directly proportional to it; (7) that the travel-time curve is continuous and has a continuous derivative. Slichter[187] has shown that the last condition is not absolute, since mathematical discontinuities do not exist in nature and since it may be possible to obtain sufficiently detailed information regarding the transition to render possible the construction of a modified solution or a sum of solutions. Witte[214] and Dahm[39] have shown that the depth may cease to be a one-valued function of the velocity without necessarily violating condition 5 above.

Both the Bateman-Knott and the Herglotz-Wiechert methods, therefore, fail if there is an unbridged first-order discontinuity where the velocity changes suddenly and causes a separation of the time-distance graph into two independent curves. This is the case in the outer crust and also at the outer boundary of the core. The crust does not create an insurmountable difficulty. The true velocities down to the Mohorovičić discontinuity can be calculated directly through the overlapping portions of the time-distance curves near the epicenter. The depth of focus and the structure of the epicentral district can be determined in certain favorably situated earthquakes, as was done by Hodgson[84,86] in the case of the Tango earthquake. Except in the case of plutonic earthquakes, the seismic energy is radiated into the interior

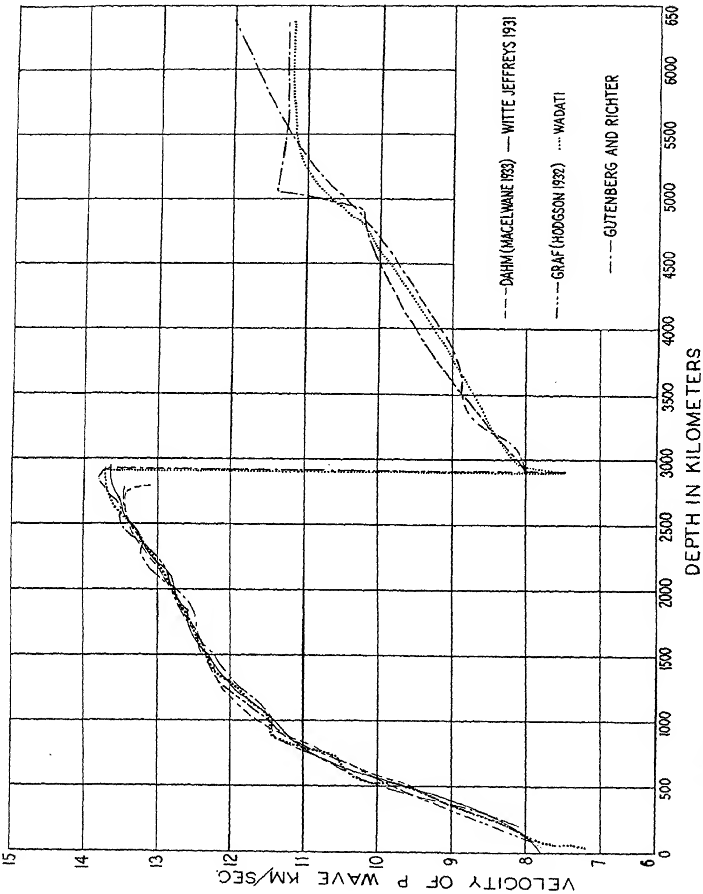

FIG. 14.—Velocity of longitudinal waves in the interior of the earth as a function of the depth.

of the earth through a comparatively small cone bounded by the critical angle of incidence on the Mohorovičić discontinuity. At many of the observing stations the crustal structure is known. Hence we may, without loss of accuracy, substitute for the true earth a hypothetical sphere stripped to the level of the Mohorovičić discontinuity and reduce the time-distance curve to that basis.

CONCLUSIONS CONCERNING THE ROCK MANTLE

In 1907 Wiechert[209,219] reported in a lecture at Dresden the results obtained by his earlier methods of calculating wave paths and velocities from the Wiechert-Zoeppritz[220] travel-time curve. He concluded that the velocity of condensational waves rises from about 8 km. per second near the earth's surface to about 13 km. per second at a depth of 1,500 km. and then drops to about 10 km. per second at greater depths. From these values he inferred that his previously proposed hypothesis of an earth composed of a stone mantle and a metal core was borne out and that the boundary between the two lay at a depth of 1,500 km. The work was remarkably ingenious for the facilities then available, though the conclusion as to the depth of the core was premature.

In the years succeeding the publication of the Wiechert-Herglotz and Bateman solutions of the integral equation of path in 1910, calculations of the relation between velocity and depth were published by Wiechert[213] and Geiger (1910), Zoeppritz,[222] Geiger and Gutenberg (1911), Geiger[53] and Gutenberg (1912), Gutenberg[55,57,60,62,65] (1914, 1923, 1927 and 1930), Mohorovičić[151] (1915), Knott[109] (1919), Wiechert (1922, unpublished), Repetti[168] (1927), Witte[214] (1932), Wadati[206] and Oki (1933), Wadati[206] and Masuda (1934), Gutenberg[71] and Richter (1935) and Dahm[39,40,41] (1936).

The degree of trustworthiness of the solutions depends, of course, on the accuracy of the travel-time curve upon which each was based. Though systematic errors underlay the older curves, these errors were balanced to some extent against each other so that the solutions did not turn out so badly as might have been expected. Among the more modern solutions of Witte, Wadati, Gutenberg and Richter and Dahm there is in the main a surprising agreement, as will be seen in Fig. 14, where they are plotted together. Witte's solution is based on Jeffreys' 1932 curve, Wadati's on his own curve and on those of Scrase and Krumbach, Gutenberg and Richter's on their own curve and Dahm's on the Macelwane curves and on his own. There has been no solution based on the Jeffreys-Bullen curves or on Jeffreys' 1936 curves. Graf's solution (Table 59) based on the Hodgson curve is included

in the figure for comparison. The numerical values are given in Tables 54 to 59.

TABLE 54
PATHS OF CONDENSATION WAVES AND VELOCITIES IN THE MANTLE
(According to Repetti[168])

$\Delta°$	d	r	v
5	29.0	6,337.0	8.01
10	54.6	6,311.4	8.06
15	136.3	6,229.7	8.18
20	292.9	6,073.1	8.55
25	512.5	5,853.5	9.26
29	719.4	5,646.6	10.06
32	897.0	5,469.0	10.83
33.2	973.0	5,393.0	11.18
35	1,000.4	5,365.6	11.28
39	1,128.0	5,238.0	11.70
39.4	1,141.0	5,225.0	11.76
42	1,174.8	5,191.2	11.81
45	1,277.2	5,098.8	12.03
50	1,414.0	4,952.0	12.36
55	1,569.6	4,796.4	12.65
60	1,723.0	4,643.0	12.90
64.5	1,855.3	4,510.7	13.11
70	1,882.1	4,483.9	13.11
75	2,004.0	4,362.0	13.12
77	2,095.1	4,270.9	13.16
80	2,135.0	4,231.0	13.17
85	2,234.0	4,132.0	12.95
90	2,254.1	4,111.9	13.11
95	2,411.5	3,954.5	13.02
100	2,584.0	3,782.0	13.03
105	2,685.0	3,681.0	

Using the term *mantle* in the Wiechert-Gutenberg sense to include all the material between the Mohorovičić discontinuity at the base of the crustal layers and the Wiechert-Gutenberg discontinuity at the outer boundary of the core, we may say that there is no general agreement as to the existence of any sudden change of velocity such as would indicate a discontinuity of the first order in the mantle. Gutenberg,[53,60,62,71,222] influenced largely by his findings on variation of amplitude with depth, has consistently placed three second-order discontinuities between the base of the crust and the top of the core. A *first-order* discontinuity is a relatively sudden transition from an upper to a lower medium, or vice versa, in which the velocities are

TABLE 55
VELOCITIES OF P AND S WAVES
(According to Witte[214])

Δ°	P				S			
	Depth, km.		Velocity, km./sec.		Depth, km.		Velocity, km./sec.	
	I	II	I	II	I	II	I	II
0	.....	......	.....	7.78	.....	......	....	4.33
2.5	.....	(5)	.....	7.79	.....	(4)	....	4.34
7.5	31	45	7.83	7.91	30	43	4.36	4.40
12.5	87	124	7.93	8.16	86	121	4.41	4.53
17.5	235	254	8.30	8.57	201	244	4.55	4.76
20.0	480	......	9.52	.....	371	......	4.87	
22.5	587	432	10.18	9.22	536	423	5.33	5.09
27.5	710	692	10.75	10.34	809	706	6.14	5.74
30.0	.....	787	.....	10.67				
32.5	834	872	11.14	11.02	898	908	6.36	6.24
37.5	886	973	11.22	11.30	918	996	6.36	6.38
42.5	1,026	1,080	11.45	11.52	999	1,067	6.40	6.45
47.5	1,265	1,251	12.01	11.86	1,178	1,203	6.54	6.57
52.5	1,345	1,404	12.10	12.15	1,323	1,294	6.66	6.63
57.5	1,474	1,518	12.26	12.32	1,429	1,360	6.72	6.64
62.5	1,678	1,618	12.56	12.40	1,558	1,529	6.79	6.71
67.5	1,695	1,750	12.56	12.52	1,578	1,698	6.78	6.79
72.5	1,896	1,957	12.66	12.71	1,779	1,829	6.83	6.84
77.5	2,195	2,195	13.00	12.99	1,975	2,008	6.90	6.91
82.5	2,400	2,417	13.22	13.25	2,196	2,238	6.98	7.01
87.5	2,586	2,574	13.44	13.41	2,598	2,418	7.24	7.08
92.5	2,734	2,643	13.56	13.44	2,746	2,573	7.31	7.13
97.5	2,858	2,682	13.62	13.41	2,817	2,715	7.34	7.15
102.5	2,936	2,682	13.62	13.41	2,891	2,766	7.32	7.15
107.5	2,936	2,682	13.62	13.41	2,943	2,766	7.30	7.15

different, thus causing a complete break in the depth-velocity curve. A *second-order* discontinuity, on the other hand, is a sudden change in the *rate* at which the velocity increases or decreases with depth, thus producing a sudden change of slope in the depth-velocity curve. In the course of the two and one-half decades during which Gutenberg has given his attention to the problem it is natural that he should have shifted the positions of these discontinuities somewhat in response to new observational data. In 1930 he placed them at depths of 1,200, 1,900 and 2,150 km., and in 1935 Gutenberg[71] and Richter placed them at about the same depths (Ref. 71, Fig. 16, page 349). Repetti in 1927

TABLE 56
VELOCITIES IN THE MANTLE OF THE EARTH
(According to Wadati and Oki[206])

$\Delta°$	P wave		S wave	
	Depth, km.	Velocity, km./sec.	Depth, km.	Velocity, km./sec.
0	0.0	5.00	0.0	2.79
2	28.3	7.21		
4	54.2	7.62		
6	77.2	7.80		
8	92.2	7.87		
10	114.4	7.94		
12	156.6	8.09		
14	217.5	8.35		
18	364.0	9.03		
20			423.1	5.20
22	526.5	9.83		
24	604.1	10.21	562.8	5.54
26	669.0	10.50		
28			713.1	5.92
30	774.7	10.91	780.6	6.07
35	889.2	11.24	887.9	6.25
40	988.0	11.44	1,004	6.41
45	1,132	11.69	1,102	6.50
47	1,221	11.86		
50	1,317	12.07	1,195	6.57
55	1,441	12.27	1,294	6.62
60	1,548	12.37	1,433	6.69
65	1,671	12.49	1,593	6.77
70	1,827	12.62	1,763	6.86
75	2,054	12.85	2,009	7.00
80	2,314	13.15	2,185	7.12
85	2,530	13.42	2,308	7.16
90	2,723	13.62	2,434	7.20
95	2,872	13.74	2,589	7.22
100	2,977	13.79	2,688	7.24

also thought he found evidence for second-order discontinuities at depths of 1,140, 1,860 and 2,100 km. The agreement is excellent; but many seismologists have doubted whether the results are not fortuitous because of the enormous scattering of the amplitude data and the differences among travel-time curves, to which the Wiechert-Herglotz method is quite sensitive.

Repetti[168] also discovered a discontinuity at a depth of about 950 km. which seemed sharp enough to cause reflections. Indeed he

TABLE 57
VELOCITIES IN THE MANTLE OF THE EARTH
(According to Gutenberg and Richter[71])

Δ, magameters	*P* wave Depth, km. − 40	*P* wave Velocity, km./sec.	*S* wave Depth, km. − 40	*S* wave Velocity, km./sec.
1.6	200	8.1	180	4.5
1.8	360	9.0		
2.0		....	380	5.0
2.4	550	10.2	550	5.5
2.8	....	....	690	5.9
3.0	720	10.8		
3.2		....	800	6.2
3.6	860	11.3		
3.8	....	....	930	6.4
4.0	920	11.4		
4.4		....	1,030	6.5
4.8	960	11.4		
5.0	1,070	11.5	1,090	6.6
5.2	1,180	11.7		
5.6	1,330	12.0		
6.0	1,400	12.1		
6.2		....	1,220	6.6
6.6	1,500	12.2	1,510	6.7
.0	.....	....	1,700	6.9
.2	1,610	12.4		
		....	1,820	7.0
	1,810	12.5		
	1,820	12.5		
7.8		....	1,910	7.1
8.0	2,050	12.9		
8.2		....	1,980	7.1
8.4	2,250	13.2		
8.8	2,340	13.2		
9.2	2,490	13.5		
9.6		....	2,100	7.0
9.8	2,580	13.5	2,150	7.0
10.0	.....	....	2,320	7.0
10.4	2,850	13.8	2,720	7.2
10.8	.....	....	2,960	7.3
11.5	2,920	13.7		

thought he was able to identify reflections from that depth on the records that he published. The work of Witte[214] and Dahm[39,40,41] corroborates the reality of the Repetti discontinuity; but their results seem to indicate that the break is probably not so sharp in reality as

TABLE 58
VELOCITIES IN THE MANTLE OF THE EARTH
(According to Dahm[40,41])

Δ° (rounded)	Macelwane *P* curve		Long Beach			
			P curve		*S* curve	
	Depth, km.	Velocity, km./sec.	Depth, km.	Velocity, km./sec.	Depth, km.	Velocity, km./sec.
0	16	7.75				
8	59	7.82				
9	.....	.....				
10	93	7.91	.....	.....	55	4.66
11	.....	.....				
12	140	8.06	57	7.94	70	4.66
13	.....	.....				
14	199	8.28	74	7.94	93	4.77
15	.....	.....				
16	267	8.56	131	7.99	128	4.49
18	350	8.94	227	8.27	156	4.51
19	.....	.....	335	8.83	195	4.56
20	433	9.33	419	9.33	308	4.76
21	.....	.....				
22	509	9.62	469	9.56	452	5.13
23	.....	.....				
24	573	9.97	.....	.....	586	5.56
26	639	10.25	611	10.21	679	5.86
28	712	10.54	.....	.....	743	6.04
30	789	10.84	775	10.98	785	6.15
31	829	11.00				
32	866	11.14				
36	981	11.54	873	11.36	844	6.24
40	1,076	11.79	947	11.55	904	6.31
44	1,156	11.95	1,023	11.70	978	6.37
48	1,236	12.08	1,101	11.83	1,077	6.46
52	1,319	12.18	1,186	11.94	1,191	6.55
56	1,408	12.28	1,286	12.06	1,294	6.63
60	1,510	12.37	1,412	12.21	1,384	6.69
64	1,621	12.47	1,533	12.36	1,469	6.72
68	1,762	12.59	1,659	12.49	1,560	6.76
72	1,925	12.76	1,765	12.59	1,670	6.79
76	2,108	12.96	1,885	12.68	1,865	6.88
80	2,316	13.22	2,055	12.83	2,052	6.99
84	2,423	13.33	2,278	13.09	2,177	7.05
88	2,523	13.39	2,531	13.44	2,324	7.11
92	2,628	13.44	2,644	13.57	2,462	7.16
95	2,678	13.45	2,680	13.58	2,539	7.18
98	2,717	13.45				
100	.....	.....				
102.5	2,780	13.42	.....	.....	2,704	7.18
			.....	.....	2,737	7.17

TABLE 59

VELOCITIES OF CONDENSATION WAVES IN THE MANTLE OF THE EARTH
(Calculated by Graf from the Hodgson P Curve of the Tango, Japan, Earthquake, Mar. 7, 1927)

$\Delta°$ (rounded)	Depth, km.	Velocity, km./sec.
10	92	7.94
12	167	8.30
14	216	8.53
16	275	8.79
18	348	9.14
20	414	9.45
24	528	9.90
28	675	10.47
32	819	11.01
36	923	11.32
40	1,023	11.58
44	1,102	11.72
48	1,271	12.08
52	1,353	12.21
56	1,410	12.27
60	1,480	12.32
64	1,628	12.44
68	1,806	12.66
72	1,910	12.76
76	1,993	12.80
80	2,256	13.07
84	2,458	13.32
88	2,560	13.41
92	2,638	13.45
96	2,705	13.45
100	2,763	13.44

Repetti's solution would make it appear. Nevertheless the Repetti discontinuity is important because it seems to be at the lower limit of the very rapid increase of velocity with depth.

CONCLUSIONS CONCERNING THE NATURE AND DEPTH OF THE TRANSITION INTO THE CORE

The phenomena of the shadow zone and of focal zones seem to leave no doubt as to the existence of a core in the earth. A number of methods have been applied to determine its size. One method uses the depth of the vertex of the grazing ray which emerges at the beginning of the shadow zone. Another uses the travel time of waves reflected from the outer boundary of the core. Still another is based on the waves that are refracted through the core. All of these depend on the

nature of the transition shell. Another difficulty arises from the impossibility of determining with the requisite precision where the curved portion of the travel-time curve for the direct rays ends and the linear portion begins because the curvature is so slight and the observations scatter and may belong to more than one curve. Furthermore, the amplitudes do not drop suddenly but fade away gradually between 90 and 110° epicentral distance. Hence, in the imperfect state of our present knowledge no two seismologists are likely to agree on the exact distance at which the grazing ray emerges or even as to whether there is such a ray in the strictest sense. Neither will all agree on the precise slope of the linear part of the travel-time curve because the points are so scattered. Besides, the very slight curvature of the later portion of the travel-time curve before it becomes linear introduces considerable uncertainties in the slope of the tangent the determination of which is essential to the Wiechert-Herglotz method. Hence the velocities in the deeper mantle become less and less trustworthy as the core is approached. Hence also arise considerable uncertainties in the calculation of the depth of reflection from the travel times of waves reflected by the core. Gutenberg[71] and Richter smoothed the travel-time curves of P, S, P_cP and S_cS from many earthquakes and further assumed a multiplicity of subparallel P and S curves. Choosing the end of the curved segment at 103.°5 they found that the grazing ray penetrated to a depth of 2,900 km., the same depth they found for the reflection of P_cP and S_cS. This is the depth at which Gutenberg[55,57] had previously placed the boundary of the core. Dahm[39,40] on the other hand determined the limit of curvature at an arcual distance of 102.°5 by finding the position and slope of a linear segment from a number of chosen observations and its point of juncture with the preceding curved segment. Now, according to his Wiechert-Herglotz solution, the ray which emerges at 102.°5 does not penetrate deeper than 2,780 km.; whereas that depth for the core boundary was totally inconsistent with his observed travel times of P_cP (Table 60); nor would any reasonable assumption as to continuous decrease of velocity agree with a depth of reflection of P_cP so small as 2,900 km. Therefore Dahm assumed a constant velocity layer as the form of transition into the core and determined the velocity and the layer thickness that would be required. He found that a sudden decrease from 13.42 km. per second to 12.57 km. per second and a thickness of 220 km. would satisfy the travel times of P_cP. The main discontinuity on this hypothesis would be pushed down to a depth of 3,000 km. (Fig. 17). Obviously this is not a unique solution. But it might suggest a modified form of the transition shell of Washington

TABLE 60

TRAVEL TIMES OF CONDENSATIONAL WAVES REFLECTED FROM THE CORE OF THE EARTH

After Dahm (Ref. 140, p. 225)		After Gutenberg and Richter (Ref. 71, p. 348)	
$\Delta°$	Travel time of P_cP	$\Delta°$	Travel time of P_cP
	m. *s.*		*m.* *s.*
10	8 47	0.0	8 38
15	8 48	13.9	8 40
20	8 52	21.0	8 52
25	9 01	28.1	9 11
30	9 15	35.2	9 29
35	9 30	41.9	9 51
40	9 46	48.0	10 11
45	10 02	56.5	10 41
50	10 18	68.0	11 25
55	10 36	78.6	12 10
60	10 54		
65	11 13		
70	11 34		
75	11 55		
80	12 16		
85	12 38		
90	13 01		
95	13 24		
100	13 47		

and Adams, or the "ground-glass" transition of Jeffreys[103] or the mechanism for the traveling reflection of Schmidt.[176] It was suggested by Wiechert and held as probable by most seismologists that the continuation of the P phase with diminished amplitudes into the shadow zone is due to diffraction around the core. Whether the phenomenon is really diffraction or whether it is scattering, or a traveling reflection, or simple refraction remains to be determined when more information of both theoretical and observational character becomes available. Wiechert[210] showed that a *sudden decrease* in the velocity as a function of depth would be a *sufficient condition* for the production of a shadow zone and a focal zone. But sudden decrease is clearly not a *necessary condition*, as has been proved, *e.g.*, by Jung,[107] who showed that a shadow zone and a focal zone must necessarily arise whether the rate of decrease is discontinuous or gradual provided that $dv/dr > v/r$, where v is the velocity and r is the radius vector from the

center of the earth, and that these phenomena could even be due to a *regular ray* whose curvature is such that it travels around a circle sending some energy back to the surface. But in every case so far considered *some type* of lower velocity *core* is required.

CONCLUSIONS CONCERNING THE CORE OF THE EARTH

Although the nature of the earth's core is as yet an open problem, certain facts have been gleaned which lead us to hope that the problem may be solved. There seems to be no reason for doubting the considerable decrease of speed of seismic waves in the central portion of the earth. It is indicated by the shadow zone and also by the small average speed of condensational waves along a diameter of the earth. The position and width of the shadow give us a rough idea at least of the size of this nucleus or central body. It is ambiguous to speak of the radius of the core unless we know for certain the nature of the boundary or, in other words, the character and thickness of the transition shell or shells. Nevertheless, the deepest level that is attained by the rays that emerge just before the shadow zone seems *certainly* to be less than halfway to the center of the earth. The depth is probably of the order of five-elevenths of the earth's radius. The shadow zone is a little more than 40° in breadth, varying in position with depth of focus but extending roughly from 102 to 143°.

At a distance of about 143° a new phase of enormous energy appears on the seismographic records with a travel time considerably greater than that of P. It can be shown to be condensational in character and was given the name P' by Angenheister.[7,8] Since the prime is otherwise used to denote rays that emerge beyond 180°, Bastings has suggested the notation PKP instead of P'. Angenheister was able to follow the travel-time curve backward from 143 to 100°. But, like the direct P beyond 102.°5, the P' or PKP waves within this range have but small amplitudes. The present writer[130] formerly thought that the suddenness of the transition from small to very large amplitudes at the outer edge of the shadow zone was good evidence for a sharp core boundary. But this can be held no longer. A study of Fig. 15, which represents Gutenberg's[57] map of condensation wave fronts and rays as seen on a central section through the earth, will reveal that the waves refracted through the core may arrive at any point more distant from the origin than 145° by two distinct paths and at two different times. The travel-time curve will, therefore, consist of two branches which coincide at about 143°. That the travel-time curve would consist of two branches was predicted by Gutenberg,[55] and the travel times for both branches were calculated and published by him in 1914.

He set up 28 separate hypotheses as to the distribution of velocities in the core and chose that distribution which seemed to him most nearly to satisfy the observed travel times. The fact that Gutenberg's calculations and the predictions based on them have stimulated and

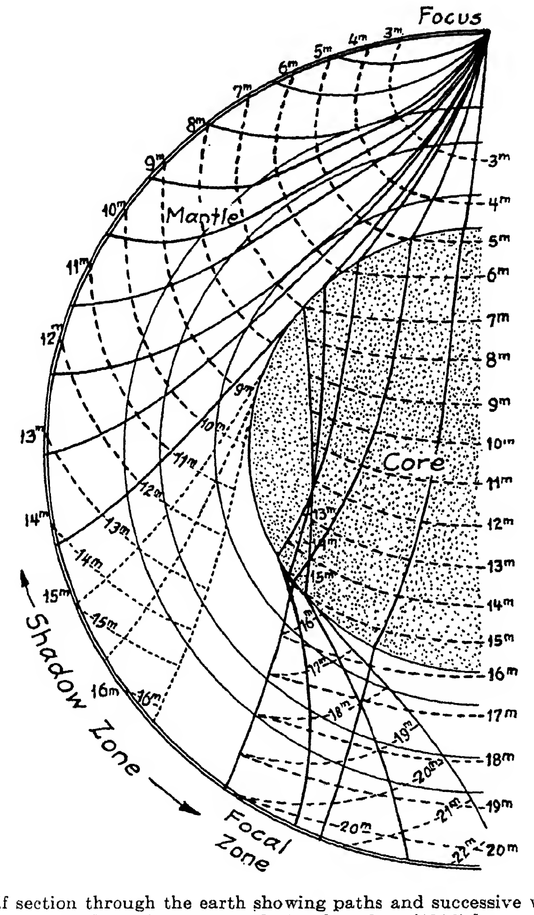

FIG. 15.—Half section through the earth showing paths and successive wave fronts of condensation waves. [*After Gutenberg* (1926).]

guided so much subsequent research on the core of the earth marks his 1914 paper as an outstanding achievement. The two branches of the P' curve were observed independently by Lehmann[124] and the present writer[131] in 1930. But there was not known even then any method for

the use of these curves in a direct attack on the structure of the core. For there would be two branches in any case, even if the slower velocity were constant all the way to the center of the earth.

Wadati (Ref. 206, **8**: 188) and Masuda have the distinction of having been the first to suggest a method for the application of the Wiechert-Herglotz procedure to the core. They pointed out that, for

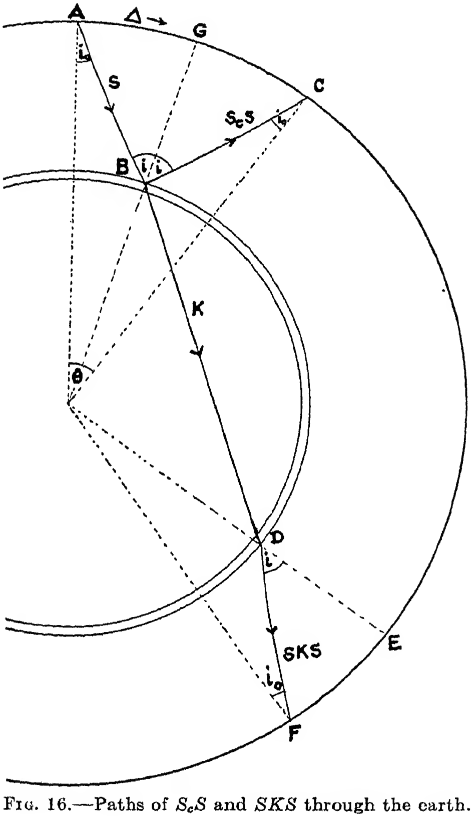

FIG. 16.—Paths of S_cS and SKS through the earth.

example, S_cS and SKS are parts of the same system and can be used to work out a travel-time curve for the surface of the core, subject, of course, to the uncertainties that surround the depth and character of the transition shell. Once an observed travel-time curve for the core is in our possession, we can apply the Wiechert-Herglotz method to find the distribution of velocities within the core. Thus, referring

to Fig. 16, if a shear wave S leaves A at the angle i_0 and impinges on the core at an angle of incidence i, a part of the energy will be reflected at B as S_cS at the same angle i and will return to the surface at C at the original angle of incidence i_0. Another part of the energy will be refracted through the core from B to D as a condensation wave. Now, the angle of refraction is unknown, and the curvature of the leg BD is unknown; but *whatever their value* the shear energy may be presumed to depart from the core at D, because of symmetry, at the same angle i at which it entered it and therefore will reach the surface at F at the same angle of incidence i_0 as that at which it left A. Thus the angle i_0 is characteristic of the ray whether as S_cS or as SKS. But i_0 is a function of $dt/d\theta$ and hence of $dt/d\Delta$. Therefore to pair off corresponding values of S_cS and SKS it is necessary only to find the value of $\Delta°$ for which S_cS and SKS have the same $dt/d\Delta$. Then the arcs $BD° = GE° = AF° - AG° - EF° = AF° - AC.°$ That is, we obtain the distance on the core surface that corresponds to the travel time from B to D by subtracting the distance of S_cS from that of the corresponding SKS; and we find the travel time for BD by subtracting from SKS that for $AB + DF = AB + BC = S_cS$. Analogous reasoning applies to each pair of every system of derived waves. The problem is then one of an earth stripped to the core. Table 61 contains the values of the velocity as a function of the radius computed in this way by Wadati (Ref. 206, **8**: 192) and Masuda and also those computed by Gutenberg (Ref. 71, **45**: 355) and Richter using the same method and their own more recent and reliable observations.

The case for shear waves in the core is much more difficult because, unlike the condensation waves that arrive at the beginning of the seismographic record, these waves will be superposed upon other radiation or will arrive simultaneously with other more prominent phases. Furthermore, the limitations that the boundary conditions place on shear waves entering and especially emerging from the core are much narrower than is the case with condensation waves. In his study of the south Pacific earthquake of 1924, the present writer[131] found two phases on many records that might be identified with the two branches of the S' phase or shear wave refracted through the core, which are analogous to the two branches of the condensation or P' wave. Bastings[11,12] found what he identified as the first branch on the records of the Buller (New Zealand) earthquake of 1929 at 16 stations and the second branch at 26 stations. He has suggested the symbol Z to represent a leg of the path of a shear wave within the core, to correspond to K for a condensational wave; so that S' would be written SZS. Imamura[89] also observed one occurrence of S' or SZS at Tokyo

TABLE 61

VELOCITIES OF CONDENSATION WAVES IN THE CORE OF THE EARTH
(According to Wadati[206] and Masuda and to Gutenberg and Richter[71,73a,73b])

Radius from the center of the earth, km.	Wadati and Masuda	Gutenberg and Richter			
	Velocity, km./sec., 1933	Velocity, km./sec. 1938	Velocity, km./sec. 1935	Depth, km.	Angular distances, θ° at the core boundary, 1935
0	11.2	11.3	12	6,370	180°
500	11.2	11.3			
580	....	......	11.5	5,800	150°
1,000	11.1	11.4			
1,350	....	......	10.5	5,010	120°
1,500	10.4	10.2			
1,680	....	......	10.0	4,690	105°
2,000	9.7	9.9			
2,130	....	......	9.5	4,240	90°
2,500	9.1	9.4			
2,550	....	......	9.0	3,820	75°
2,810	....	......	8.9	3,550	60°
3,000	8.6	8.7			
3,010	....	......	8.7	3,350	41°
3,200	....	......	8.3	3,170	30°
3,400	8.1	8.0±			
3,440	7.9				
3,446	....	8.0±	7.5–8	2,920	0°
3,457	7.5 inside 13.8 outside	13.7			

in the case of the south Atlantic earthquake of June 27, 1929. In view of the disagreement of other competent seismologists, the inherent difficulties of observation and interpretation and the incomplete and unsatisfactory state of mathematical wave theory and of our knowledge of waves in general, the final and definite decision as to the reality of shear waves in the core must be left to the future. But in the meantime seismological evidence must not be discounted or rejected because of presumptively contrary evidence from other fields, which on careful analysis is found to be subject to even greater uncertainties.

The picture of the interior of the earth (Fig. 17) which thus results from our evaluation of the seismological evidence is a provisional and tentative one, blurred in many of its lines, sketchy in its details. Yet we seem to have evidence that will satisfy the most skeptical, (1) that

there are crustal layers in many places; (2) that there is an outer region or shell some 600 miles in thickness in which, at least in the gross, the speed of elastic body waves and therefore the ratio of the elasticity to the density increases rapidly with depth; (3) that there is an intermediate region or shell in which the rate of increase of speed is more doubtful (though in general it is much less and may go over into a

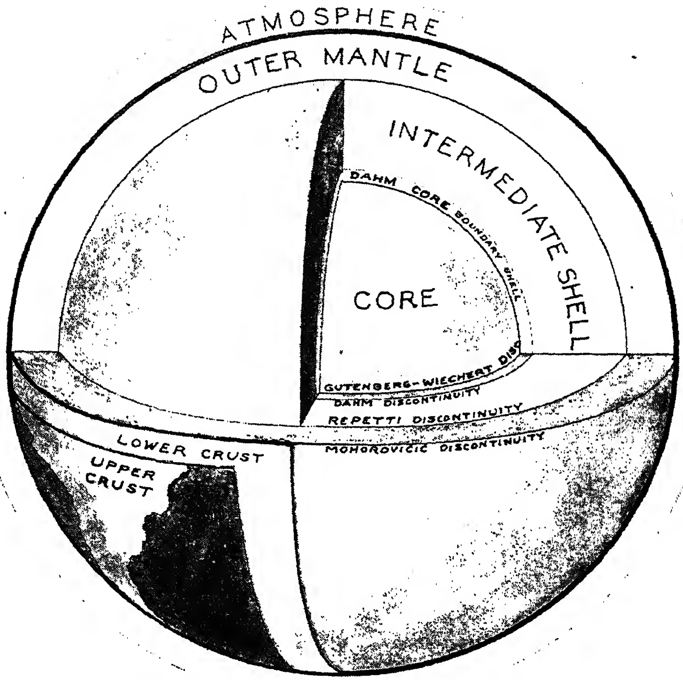

FIG. 17.—The structure of the earth. (*After J. B. Macelwane.*)

decrease with depth) but in which nevertheless the ratio of the elasticity to the density is many times higher than any of which we have experience on the surface of the earth; (4) that there is at the center of the earth a nucleus or core which strongly reflects and refracts elastic waves. The diameter of this core probably measures somewhere between ten- and eleven-nineteenths that of the whole earth. The average velocity with which condensation waves are transmitted

through this core, though much less than in the intermediate shell, is still nearly twice that of sound in steel in the laboratory. The speeds in the core would seem to be least near the periphery and to increase considerably toward the center.

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CHAPTER XI

EVIDENCE FROM DEEP-FOCUS EARTHQUAKES

B. GUTENBERG
(*Professor of Geophysics, California Institute of Technology, Pasadena*)

AND C. F. RICHTER
(*Assistant Professor of Seismology, California Institute of Technology, Pasadena*)

There is no longer any reasonable doubt that earthquakes originate at levels ranging from the surface down to depths of at least 700 km. Recent investigation tends to support a classification into normal, intermediate and deep earthquakes; the boundaries between these classes are not easy to define in a perfectly satisfactory way.

Normal earthquakes, which appear to constitute a large majority of recorded shocks, originate comparatively near the surface. Of these practically all the largest, as well as a great number of smaller ones, are tectonic in character, being often demonstrably associated with faulting; others are due to volcanic activity or to minor causes such as the collapse of caverns.

Intermediate earthquakes were originally defined by Wadati[1] as those originating at depths between 100 and 300 km. Most of the shocks in this range show a characteristic geographical distribution that distinguishes them both from normal shocks and from the true deep-focus shocks (see Fig. 18). The present authors have usually assigned shocks from 60 to 250 km. to this group, but the limits are not definitely fixed and may vary regionally.

GEOGRAPHICAL DISTRIBUTION

The distribution of deep-focus earthquakes was first discussed by Turner[2] and has been examined since by various authors.[3] We here summarize the most recent conclusions, based on a thorough revision of epicenters and depths.[4]

Intermediate shocks occur in nearly all regions in which normal earthquakes are frequent. The greater number are associated with the circum-Pacific belt of normal seismic activity; but intermediate shocks have also been identified (to date) in Burma, the Hindu Kush, Persia, the eastern Mediterranean, Rumania, in and near Italy, and in the south Atlantic (about 59°S. 27°W.). The Hindu Kush shocks provide

a unique series of repetitions from nearly the same epicenter and depth (about 220 km.) over a period of 30 years. Recently there has been an average of one fairly strong shock from this source every year.

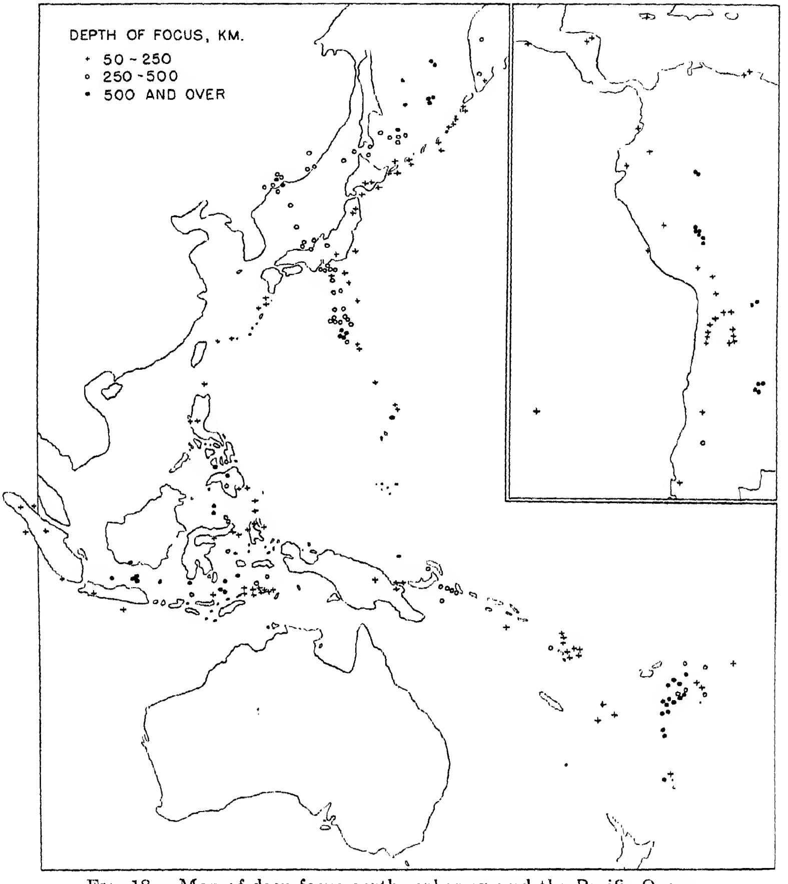

FIG. 18.—Map of deep-focus earthquakes around the Pacific Ocean.

Deep-focus earthquakes in the restricted sense are known at present only from the Pacific region. However, observations adequate for identifying such shocks are available only for a comparatively short time, not much exceeding 30 years, and are reasonably full only for the latter half of this time. Consequently, our knowledge of their distribution is necessarily incomplete.

One of the most striking results is the absence of deep or even intermediate shocks from North America proper. Intermediate shocks occur in Mexico and Central America; their occurrence in the West Indies is suspected. The Canadian earthquakes of 1925 and 1935 appear to have originated at slightly greater depth than that of the average normal shock; but it is doubtful whether they should be considered even as intermediate shocks. This also applies to a number of shocks in the Aleutian Islands and Alaska.

In South America there are several groups of shocks at depths from 600 to 660 km.; these epicenters lie chiefly east of the Andes, with which they have no apparent connection. It is possible that the epicenters lie on a continuous active zone, but this cannot be considered as established. Careful examinations have thus far failed to confirm the occurrence in South America of any shocks at depths between 290 and 600 km. There is an equally evident gap in the geographical distribution, as the rather frequent intermediate shocks (at depths from 100 to 290 km.) are all in or close to the Andean zone. Most of the stronger normal shocks of South America originate still farther west, being particularly frequent in the oceanic deeps and the coastal plain.

The activity in the Japanese area is exceptionally well known, owing to the large number of observing stations and to the detailed investigations by Wadati and others. The geographical relations are largely analogous to those in South America. Though the principal seismic activity at normal depth is associated with the deeps lying off the Pacific coast, the epicenters of intermediate shocks lie farther inland. There is one zone extending from the Marianas Islands to eastern Honshu and along the Kuril Islands to Kamchatka, and another zone in the region of Kyushu and the Ryukyu Islands, apparently continuing by way of Formosa to the Philippines. As appears from Fig. 18, the true deep-focus shocks lie still farther "inland" (*i.e.*, still farther from the Pacific basin); they occur along two principal zones, which are almost rectilinear, and intersect nearly at right angles. One lies west of the Marianas and Bonin Islands, extending transversely across central Japan and the Japan Sea to the region of Vladivostok; the other extends northeasterly from Vladivostok toward Kamchatka. The distribution in depth is unlike that in South America; the depths range almost continuously from 300 to 650 km., with the deepest foci occurring at the ends and at the intersection of the two zones.

The region of the Dutch East Indies and Philippine Islands presents a very complicated geographical distribution of deep shocks. The very deep shocks are again inland from the margin of the Asiatic

continental area; in the Philippines this means that they occur within the island area, far west of the region of normal seismic activity in the deeps off the Pacific coast. There is a belt of very deep shocks extending through the Java Sea and Flores Sea; the corresponding intermediate shocks occur under Java and Sumatra, and the normal shocks are still farther out, along the margin of the Indian Ocean.

A line of intermediate shocks extends from northern Celebes to Halmahera. This rather exceptional trend is followed here by known tectonic lines, recent volcanism and gravity anomalies.

In and about the Banda Sea, shocks of all types occur, from normal shocks down to depths of 600 km.

Intermediate activity extends from New Guinea through the Solomon Islands and New Hebrides to the Loyalty Islands. In the Solomon Islands and Bismarck Islands shocks also occur at about 400 km.

The area included between the Fiji Islands and the Tonga and Kermadec deeps is one of great activity at almost all depths. Normal shocks, some of them very large, occur chiefly to the eastward, near the deeps. Intermediate shocks appear to occur slightly farther west, but in this region it is difficult to distinguish them from normal shocks. Deep shocks occur still farther west, the very deepest (down to 680 km.) occurring about 20°S., 180°.

The present authors have not been able to verify the occurrence of intermediate or deep shocks in the vicinity of New Zealand from data thus far available, although the occurrence of intermediate shocks there is by no means improbable.

The South American region of activity appears to be completely isolated from the active areas of the western and southwestern Pacific; there is no evidence of deep-focus shocks in the southeastern Pacific or in the Antarctic which might suggest a connection between the two regions, although normal shocks are of occasional occurrence there.

CORRELATION WITH OTHER PHENOMENA

Intermediate shocks are conspicuously related to the surface geology. Their epicenters invariably fall on or close to tectonic lines of Tertiary or more recent mountain building. Consequently, they frequently run parallel to the lines of epicenters of normal shocks. In the Pacific region the relation is often such that the normal shocks occur on the edge of the continental shelf where it slopes down into a foredeep, while the epicenters of the intermediate shocks are within the continental area, often along the axis of a chain of islands or a mountain

range. As a result of this, it often happens that the epicenters of intermediate shocks follow a line of active or recent volcanism.

As pointed out by Wolff,[11] the present volcanic zones originated in the Tertiary, and consequently their position corresponds to the forces active at that remote epoch and does not coincide with the present zones of tectonic activity. Thus it appears probable that contemporary intermediate shocks and volcanic activity are both due to the same remote cause, but it is unlikely that there is any direct causal connection between these shocks at depths of as much as 250 km. and the volcanic activity at or near the surface.

In the East Indies, in the Japanese region and probably in the West Indies, the lines of large gravity anomalies correspond better with the normal shocks than with the intermediate shocks. For other regions the gravity data are not sufficient.

The distribution of really deep foci shows no direct relationship with the surface geology. The only evidence of such a relationship is that most of the epicenters occur in zones which usually run parallel to the larger structures, and lie often as much as several hundred kilometers inland from the zones of normal and intermediate activity around the Pacific basin.

MECHANISM

Any theory of the structure and physical state of the interior of the earth, outside the central core, must be consistent with what is known and with what theories are held as to the occurrence and mechanism of production of deep-focus earthquakes. The present writers have repeatedly[4,5,6] expressed the opinion that the causes of these earthquakes are in no essential respect different from those of shocks nearer the surface.

Available data show no clear gap in depth separating the various types of shocks, when the shocks occurring in all parts of the world are taken together. Apparently normal shocks are definitely most numerous; the frequency decreases rather slowly with increasing depth, so that intermediate shocks occur with about equal frequency through their range. The transition between normal and intermediate shocks has not yet been investigated in sufficient detail. The seismograms of shocks at these depths are unusually difficult to interpret, so that in the present state of knowledge and with the ordinary accuracy of the data it is often impossible to assign depths with sufficient precision. At great depths the statistical tabulations show no preference for any particular depth (above the 700-km. level); however, this is not the case for the individual regions, since nearly every active zone has one or

more characteristic depths about which the shocks are more numerous. It is especially striking that no earthquake has been found to originate at a depth much in excess of 700 km., although some of the very deepest shocks are among the largest recorded, judging from the amplitudes of their seismograms.

Bullen[12] has recently correlated this fact with the increase in electrical conductivity which appears to take place at about 700 km. (see Chap. VII), with the rapid increase in velocity in seismic waves, (see Chap. X) and with his own findings of a rapid increase in density. He concludes that "a number of distinct lines of evidence are in good accord in suggesting a change in properties at a depth of order 500–700 km. below the earth's surface."

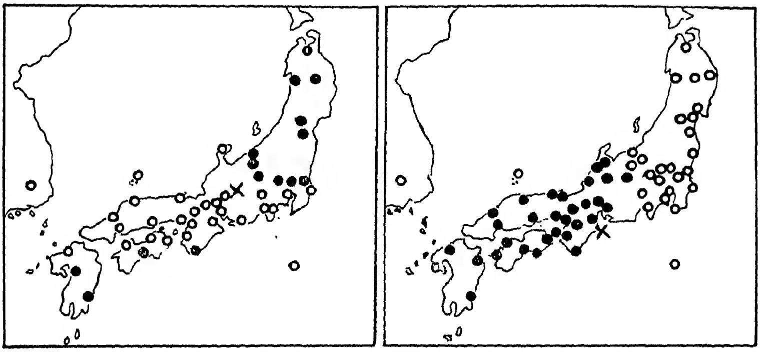

FIG. 19.—Directions of initial motion (solid circles, from the focus; open circles, towards the focus) in Japanese earthquakes. Left, June 2, 1931, depth about 260 km; right, June 2, 1929, depth about 360 km. Epicenters marked by crosses. (*After M. Ishimoto.*[7])

The large shear waves observed in deep-focus shocks leave no doubt that the causative mechanism cannot be of an explosive character but, as in the case of normal shocks, must depend on the release of strains. This is further confirmed by the pattern of initial compressions and dilatations as observed at stations in various azimuths and at varying epicentral distance[7] (Fig. 19). The data exhibit zones and sectors of initial compression and dilatation, separated by nodal lines, as would be expected when the source is complex. Similar phenomena have been observed in normal shocks, though usually less clearly indicated by the observations; this is due to the fact that the first motion of a normal shock, as recorded by a distant seismograph, is usually less sharp than that of a deep shock.

Observations at Pasadena as well as at other stations show a consistent recording of either initial compression or initial dilatation

from shocks in a given region, over long periods of time.[4] It is highly remarkable that this result is independent of focal depth. For example, shocks in western South America regularly record at Pasadena with initial dilatation; this applies to shocks at depths of about 600 km., to intermediate shocks at about 200 km. and, with rare exceptions, to normal shocks.

Further investigation is called for; however, the data are already sufficient to indicate that deep-focus shocks as well as normal (tectonic) earthquakes originate in a shearing or faulting movement which has the same direction over large areas, persists over long periods of time and is frequently the same for all focal depths in a given region. This is consistent with the geological evidence of progressive similar displacements along parallel or nearly parallel faults over extended areas.

Additional evidence for uniformity in mechanism of origin for normal and deep shocks is provided by the occurrence of aftershocks of the larger deep earthquakes. Such aftershocks are often overlooked; but careful examination of reports and original seismograms leads to the conclusion that such aftershocks are probably not significantly less frequent than those of the larger normal earthquakes.

Apart from aftershocks, there are also repetitions of shocks from nearly the same deep focus within a comparatively short time (a few months or years). The very frequent Hindu Kush shocks have already been mentioned; they plainly show that a rapid accumulation of large strains at a depth of 220 km. is quite possible.

The occurrence of shearing displacements down to depths of the order of 700 km. gives rise to no difficulty on principle; there is no necessity for great strength at these depths, since, as Haskell[8] has shown, the viscosity is so high that rapid flow would not occur even if there were no strength (see Chap. XV). This removes any difficulty in reconciling the requirements of isostasy with the occurrence of deep-focus shocks. For the existence of isostasy and for the occurrence of gradual movements of adjustment in the earth's crust, the assumption of small strength at large depth seems to be necessary; but in spite of this small strength, rupture may occur as a consequence of a comparatively rapid accumulation of large strains. This is well supported by the results of recent experiments at high pressures. Thus, Griggs[9] states: "Contrary to common belief, experiments show that when a rock enters the region of plastic flow, it will not deform indefinitely, but will rupture if the deformation is carried far enough." Bridgman[10] reports that violent and spasmodic snapping and jumping, indicating the occurrence of internal rupture, may be incidental to plastic flow.

CONCLUSIONS

Normal (tectonic) earthquakes indicate the regions of present tectonic activity. Intermediate shocks, however, are chiefly associated with lines of less recent tectonic activity (generally Tertiary). The true deep-focus shocks appear to be associated with the boundary of the Pacific basin, from which their epicenters are usually several hundred kilometers distant (inland). Thus they seem to be associated with events that took place very early in the history of the earth. Although the present writers do not wish to offer this as a serious hypothesis, it is natural to suggest that there has been motion over a long period of geological time, by which the uppermost layers surrounding the Pacific basin have been displaced toward its center relative to the lower layers, and that no new zones of faulting or weakness have developed at great depth. Nevertheless, it is an important fact of observation that displacements are still occurring at great depths around the Pacific basin.

The repeated motion in the same direction for both normal and deep shocks in given regions suggests a motion of each continental block as a whole—a motion that may be a rotation, a translation or both.

There is no evidence of a special causative mechanism for the production of deep shocks; the forces acting appear to be identical with those which occasion earthquakes at normal depth. Whereas at normal depths the accumulation of strain is made possible by the strength of the rocks, at the greater depths the high coefficient of viscosity is sufficient, and no conclusion as to strength can be drawn.

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CHAPTER XII

STRUCTURE OF THE CRUST. CONTINENTS AND OCEANS

B. Gutenberg

(*Professor of Geophysics, California Institute of Technology, Pasadena*)

and C. F. Richter

(*Assistant Professor of Seismology, California Institute of Technology, Pasadena*)

It is well known that the superficial topographic distribution of land and water over the globe is a partial but imperfect index of the extent of important structural units. This is shown by the data of the following table, which gives the areal extent of the land or sea bottom lying between the specified levels.[1]

TABLE 62
Areas at Various Levels
(According to Kossinna)

Km.	Area (10^6 sq. km.)	%	Km.	Area (10^6 sq. km.)	%
Above 5	0.5	0.1	0 –0.2	28.3 } 43.7	8.5
4–5	2.2	0.4	0.2–1	15.4 }	
3–4	5.8	1.1	1 –2	15.2	3.0
2–3	11.2	2.2	2 –3	24.4	4.8
1–2	22.6	4.5	3 –4	70.8	13.9
0.5–1	28.9 }		4 –5	119.1	23.3
0.2–0.5	39.9 } 105.8	20.8	5 –6	83.7	16.4
0 –0.2	37.0 }		Below 6	5.0	1.0

Levels are given in terms of kilometers above or below mean sea level, and areas in millions of square kilometers. The table plainly shows two principal maxima of frequency, one near sea level and the other about 5 km. below it.

Such a study of levels at once enables us to distinguish the true oceans from the shallow continental seas, under which the structures are of the same type as those of the continent. Further study and investigation separate the oceans into two groups and end by dividing the whole surface of the earth into two regions, distinct in structure and in geological history; one the Pacific basin, the other including the present continents and continental seas, together with the Atlantic

and Indian oceans. The second, or "continental" region, consists largely of areas that are known or inferred to have been land and marine at different periods.

GEOLOGICAL EVIDENCE*

The distinction between Atlantic and Pacific structure was emphasized by Suess in 1888.[2] He considered the Atlantic structure as characterizing not only the coasts of the Atlantic but also those of the Indian Ocean. These coasts are typically broken, whereas coasts of the Pacific region are typically smooth and curving, being determined by the trend of folded mountain chains.

Suess was also among the first to recognize the existence of a line of demarcation in the western Pacific region, separating structures of two different types. Born[3] refers to this as the *andesite line*, since on the western side of this line the younger eruptive magmas are principally andesitic and on the eastern side predominantly basaltic (see Fig. 20). The exact relation between the eruptive rocks and the structure is not clear. We regard this line as the true boundary that separates the Asiatic and Australasian continental area from the Pacific basin. The island arcs lying west of this line have the characteristic structure of folded continental mountain ranges, whereas the islands to the east, in the Pacific basin, appear to be free of any association with folding or orogeny, consisting of volcanic peaks, either isolated or in chains.

To the north, the boundary of the Pacific basin is easily followed along the Japanese Islands, Kamchatka and the Aleutian Islands, to the coast of North America. Although the principal boundary apparently continues south along the coasts of Central and South America, the separation of structures is not so definite as in the western Pacific. Much evidence, chiefly seismic (see the discussion later in this chapter), points to the existence in the southeastern Pacific of isolated but possibly very extensive areas of continental structure.

The much discussed rocks of Easter Island have been reported on in a recent paper by Mark C. Bandy.[34] The island is exclusively volcanic. Though Bandy finds andesites and other related rocks, he regards them as the result of normal differentiation of basaltic magma and rejects them as evidence for an underlying continental structure.

The southern boundary of the Pacific area is not known. The steep gradients descending from Marie Byrd Land into the Antarctic basin may be part of it.[31] There are a few outlying isolated areas that appear to have the Pacific type of structure. A possible case of this

* See also Chap. III.

kind is in the West Indies, where most tectonic charts show the Pacific coastal structures of North and South America connected by a loop through the Antilles (see Fig. 22). This area is cut off from the present Pacific by the comparatively young volcanic region of Panama.

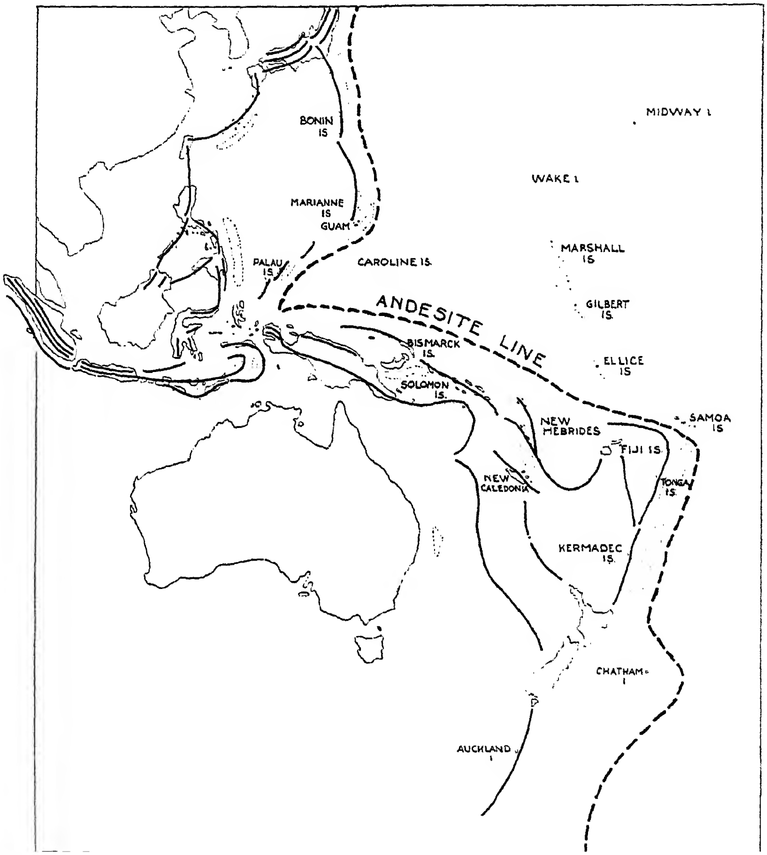

Fig. 20.—Structural map of the southwestern part of the Pacific Ocean. (*After A. Born.*)

An analogous loop, apparently without barrier against the Pacific basin, extends from South America through the Falkland Islands and South Georgia, and returns westward in Antarctica. A basin in the Arctic Ocean north of America, although completely separated from the Pacific by a wide continental area, is characterized by great depths

close to the coast.[4] The seismic evidence suggests a limited area of Pacific or similar structure.

As already pointed out, the area of the Atlantic or continental structure includes the Atlantic and Indian oceans as well as the present continents and continental seas. It is important to inquire how far the structure over this whole area is uniform. There is no doubt that structures found on the continents extend out into these oceanic basins; this is supported by the geology of islands and more recently by geological and geophysical work near the Atlantic coast of America.[5,6] However, the geophysical results[6] have been adversely criticized, and further confirmation is needed. It is possible that the Atlantic and Indian ocean basins actually differ in structure from the present continents, because of a difference in geological history. Practically all of the present land areas have been submerged at one period or another, but only under generally shallow continental seas, which could have approached the depths of the present ocean basins only in comparatively narrow geosynclinal depressions. On the other hand, all theories that do not postulate continental displacement call for the past existence of land bridges extending across the location of the present oceans.

EVIDENCE FROM GRAVITY DETERMINATIONS

The measurement of gravity on the surface of the earth gives information about the variation of density, or distribution of mass, in the interior. Theoretically, to any given surface distribution of gravity there may correspond any one of an infinity of internal distributions of mass; however, these theoretical possibilities are limited in such a way that it is often possible to draw definite or highly probable conclusions from the data.

The most general conclusion from gravity determinations is the existence of isostasy.[7] This is an expression for the fact that the masses in the crust tend to be distributed so that in all columns of a given cross-sectional area, the load above a certain fixed depth is constant. The cross section must not be taken too small, as in every region there are relatively small areas that are not isostatically in equilibrium. Cross sections of about 10,000 sq. km. (radius of the order of 60 km.) are usually sufficient to exhibit isostasy.

The depth of isostatic compensation depends on the strength of the rocks and seems to differ somewhat in different regions. The numerical values calculated for this depth depend on the assumptions made. Heiskanen[8] has found that isostatic compensation at depths

of the order of 50 km. represents the observed gravity distribution very well, on the assumption that the density distribution corresponds to that suggested by seismological investigation on the thickness of the various crustal layers. With the considerably different assumptions used in the work of the U. S. Coast and Geodetic Survey, the depth of compensation is found to be about 100 km.[7]

Theoretically, there is no reason why the depth of compensation should be marked by any sudden change in physical properties, and there is no direct evidence indicating anything of the kind. It probably represents a depth at which the strength of the rocks begins to decrease. Below this depth there can be no significant differences in density horizontally.

The processes by which isostatic equilibrium is maintained must be extremely slow, and consequently this equilibrium is liable to disturbance by geological events. However, there are certain changes that take place slowly enough so that isostatic readjustment keeps pace with them. Others take place so rapidly compared with the speed of isostatic adjustment that large gravity anomalies make their appearance. There is an intermediate class of processes, which are slow enough to allow a partial adjustment.

Examples of the first category are provided by the building up of large deltas, resulting in thick sedimentary deposits; for the geological data show plainly that the rock floor under the delta subsides gradually under the increasing load, with the result that the new sediments are constantly being laid down in shallow water. This is confirmed by the few gravity observations available in delta areas, which show plainly that there are no gravity anomalies in such areas so large as would be expected if the addition of the sedimentary load were not compensated isostatically.[9]

A probable instance of intermediate character is the lag in complete compensation of the load provided by the continental ice sheets of the Pleistocene. This is shown by the recoil of the tracts unloaded by the melting of the ice. Although the gravity anomalies now remaining in the regions affected are not known to be systematically negative or systematically different from those found in similar but unglaciated areas, there are continuous motions of uplift, both in the Scandinavian area (Ref. 10. Ref. 8, page 941) and about the Great Lakes.[35,11] These upwarpings are usually interpreted as recovery toward equilibrium, following removal of the ice load. Haskell[12] has given a theoretical discussion to show that the chief factor retarding isostatic adjustments of this kind is the high viscosity in the interior of the earth.

If geological events occur rapidly, isostatic adjustment long remains incomplete. Thus, comparatively large anomalies are found in the younger mountain ranges and over many deep oceanic troughs (where anomalies of -200 milligals are not unusual) and are particularly noteworthy where present tectonic activity is high, as in the

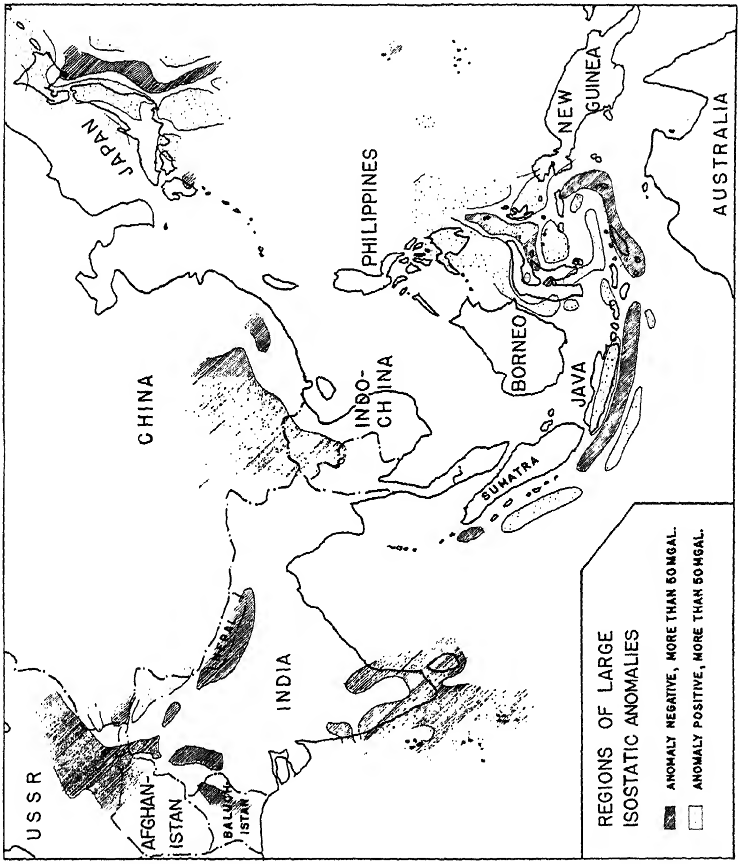

FIG. 21.—Regions of large isostatic gravity anomalies in southeastern Asia. Note that in large areas the anomalies are not known.

East Indies. Vening Meinesz[13] and others (Chap. IX) have made use of the gravity data for the interpretation of geological processes.

The numerous gravity determinations now available clearly demonstrate the greater density of the rocks in the Pacific basin area, as compared with the rocks of the continental crust. The phenomenon

of isostasy is consistent with the existence of heavier rocks beneath the continental rocks, which are supported by them as a lighter solid mass is supported in a heavier liquid. However, the mass that is thus supported includes the upper part of the heavier rocks (sima) as well as the lighter rocks (sial) of the continental structure.

Gravity surveys have been used extensively for the investigation of local structure. The conclusions are usually less definite than those reached by seismic methods; nevertheless, valuable data are often obtained with reference to the thickness of the sedimentary layers and the depth of the Basement Complex. Examples of this will be found in the literature referred to[8,10,16] (see also Chap. III).

Figure 21 is a tentative map of the known areas of large gravity anomalies in and about eastern Asia. For the East Indies and Philippines the data are those of Vening Meinesz;[13] for the other areas shown various sources have been used.[20] Areas with large positive anomalies are dotted; areas with large negative anomalies are cross-hatched. The various authors consulted have used different methods of reduction, so that their results are not exactly comparable; however, we have tried to indicate those areas where the isostatically reduced anomaly exceeds 50 milligals. There are important large regions, such as that between Japan and the Philippines, for which no data are available. The anomalous areas are bounded by solid lines where the data are definite, but the boundaries are left dotted where the results are more or less doubtful.

Figure 22 is a similar map for the West Indian area. It is based on data reported by Vening Meinesz,[36] Ewing[37] and Hess.[38] A map by Hess has been used as base and as source for much of the data, particularly the serpentinite intrusions which are indicated by crosses. Hess shows these intrusions occurring still farther south toward the western coast of South America.

The large negative anomalies in this area clearly occur along a definite structural belt; but they are not continuous throughout its length. Hess is of the opinion that the occurrence of the serpentinite intrusions indicates the course of this belt, which may formerly have been characterized by negative anomalies that have since returned toward isostatic equilibrium. Hess accordingly extends the structural belt to pass north of Cuba and thence into Central America.

The general conditions shown in the two figures are strikingly similar; in both cases we have a narrow strip in which large negative anomalies occur, usually accompanied by positive anomalies on both sides of the strip. There is as yet no general agreement as to the interpretation of these facts (see Chap. IX).

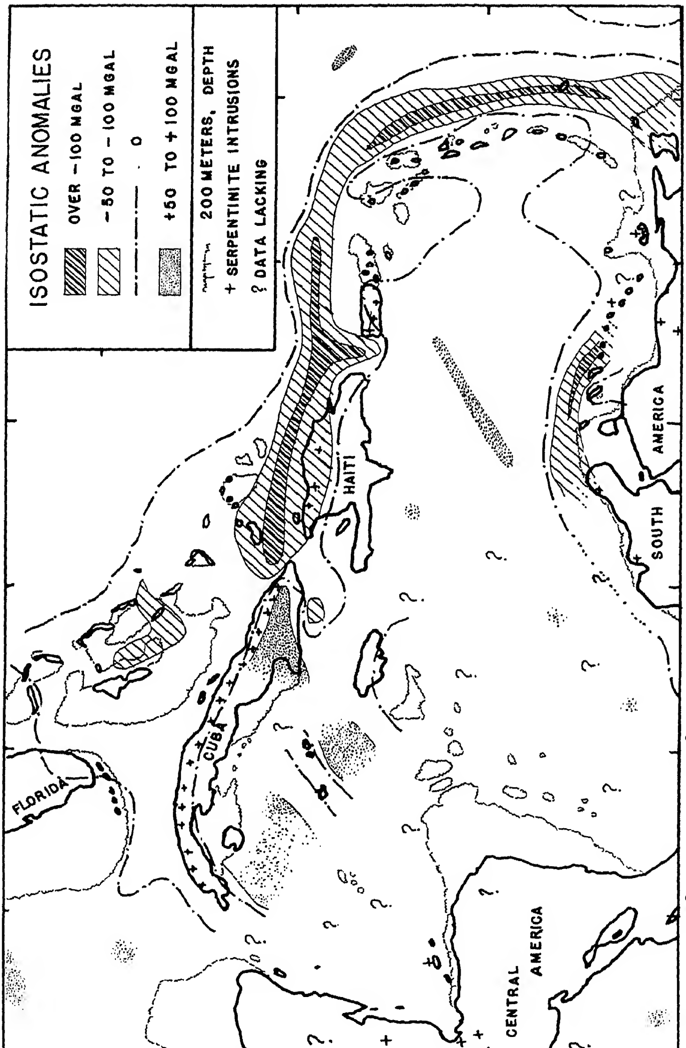

FIG. 22.—Regions of large isostatic gravity anomalies in Central America.

THERMAL AND MAGNETIC EVIDENCE

The observed thermal conditions in the crust of the earth have been discussed in Chap. VI. These observations give some information about the structure of the continental crust, but no comparable data are available for oceanic areas. Pekeris[14] has given a theoretical discussion of the effect of differences in temperature in the interior, which, he shows, under certain conditions may set up thermal convection sufficient to produce very great stresses. The maximum stress difference occurs at the bottom of the crust over the center of the oceans or continents.

Data are now available on the geographical distribution of various magnetic elements.[15] Most of these distribution patterns are peculiar and bear no evident relation to the large structures discussed in the present chapter. However, Fleming[40] concludes that a world-map of secular-variation activity for the interval 1885 to 1922 "shows again the fact brought out by Fisk, that most secular-change activity occurs in that portion of the Earth which is assumed to have a granitic crustal layer; the Pacific Ocean is remarkably free from secular-change activity." (See also Ref. 41.)

Detailed magnetic surveys in continental areas occasionally yield useful information as to the local structure of the upper part of the crust, chiefly of the sedimentary layers.[32]

EVIDENCE FROM SEISMIC DATA

The most important methods by which information about the interior of the earth can be derived from seismic data are described in Chap. X. Since we are now particularly concerned with the structure of the crust, certain additional points are here introduced or given more detailed discussion.

The geographical distribution of epicenters is a partial indication of the distribution of active faults. The epicenters of deep-focus earthquakes have been discussed in Chap. XI; in the present section, we are chiefly concerned with normal shocks, for they give information about the condition of the crust of the earth.

Most of the seismic activity of the globe occurs in the two principal belts of activity; one, the more active of the two, surrounds the Pacific, and the other extends through the Mediterranean area of Europe eastward through Asia to join the Pacific belt in the East Indies.[17]

The Pacific belt is simple in its northern and eastern section; it can be traced as a single and almost continuous line of epicenters from the east coast of Japan by way of Kamchatka, the Aleutian Islands and the

coasts of the two Americas. The principal activity is off the coasts, often following the oceanic deeps. The arc in western North America is exceptional; from Alaska to northern Mexico there are no considerable deeps off the coast. The larger shocks often take place far inland in the Californian area. From British Columbia to northern California the principal shocks, although they often occur far off the coast, are probably still within the margin of the continental shelf.

As indicated, the main belt of activity passes directly down the coast of Central America, extending by way of the Gulf of Panama to Colombia; but there is a pronounced alignment of epicenters through the West Indies and northern South America, following the line of the loop of structures (see Fig. 22) that usually appears on tectonic maps. (There is no evidence that this West Indian activity forms part of the Mediterranean belt; practically no epicenters are located between the Antilles and the Mid-Atlantic Ridge.) The occasional strong shocks in the south Atlantic appear to confirm the activity of the similar tectonic loop that extends from the tip of South America by way of the Falkland Islands, South Georgia and other southern islands into Antarctica.

In central Japan the active belt splits into two branches, one of which continues southward, following the continental side of the andesite line along the western boundary of the Pacific basin (Fig. 20), whereas the other extends through southern Japan, Formosa and the Philippines to the region of the Moluccas, where the two lines approach so closely that their exact relation in this region cannot be decided from the seismic evidence alone. The eastern branch, lying everywhere on the continental side of the andesite line, can be followed through the Solomon Islands, New Hebrides, and Fiji area to the Tonga and Kermadec deeps and southward past New Zealand to the vicinity of Macquarie Island. Since almost nothing is known of seismic activity in the Antarctic (except that large shocks are infrequent), there is no evidence whether or not this line of epicenters should be connected across Antarctica with that extending from South America into the south Atlantic.

The second major belt of activity appears as a continuation of the western branch of the Pacific belt; the epicenters closely follow Vening Meinesz' belt of negative gravity anomalies from the Philippines to the islands about the Banda Sea and thence westward, off the southern coasts of Java and Sumatra. However, epicenters in the region of Celebes, the Banda Sea and the Moluccas are often subject to large uncertainties, partly due to the frequent occurrence of intermediate and deep shocks. From Sumatra the active belt extends northward

by the Nicobar and Andaman Islands into Burma, where it turns west across Asia, following the Alpide belt of structures, especially on its south side, to the western end of the Mediterranean and some distance beyond into the Atlantic. Whether it actually connects with the mid-Atlantic belt cannot be decided from the data.

The Mid-Atlantic Ridge is marked out very well by a line of epicenters extending from the equatorial region to Iceland. Although these shocks are usually not large, their epicenters are well determined, on account of the favorable location of the American and European stations. Circumstances are not so favorable in the south Atlantic; apparently the activity is less than in the north Atlantic, and the few rather doubtful epicenters leave it undecided whether the active belt should be connected with the strong activity near South Georgia (Pacific belt) or whether it extends round by the south of Africa into the Indian Ocean.

It is quite certain that there is no belt of activity connecting the West Indian area with the Mediterranean across the Atlantic, as indicated on most of the older seismic maps of the world.

It is doubtful whether the mid-Atlantic activity extends much north of Iceland; there are a few scattered epicenters of small shocks that suggest a continuation to the north and east by way of Spitzbergen along the Arctic coast of Asia possibly as far as Bering Sea. There is no doubt about the considerable minor activity in the region of the New Siberian Islands and the Nordenskjöld Sea. This has been made the subject of special study by Tams.[18]

As the region of the East African rifts shows geological evidence of great recent activity, it is noteworthy that large shocks in this region have been rare in the last 25 years. However, the historical record shows considerable evidence of activity, which is supported by instrumental recordings for several large shocks about 1910 and 1912.

In the western Indian Ocean there is an alignment of epicenters extending from the vicinity of the Chagos Islands across Mauritius to the Crozet Rise; as already mentioned, it is possible that this line should be connected south of Africa with the Atlantic belt. The activity is in general minor, except in the vicinity of 34°S. 57°E., where numerous shocks have originated in recent years, several of them quite large. The earliest shocks assigned to this epicenter in the "International Seismological Summary" occurred in 1925; but it is quite possible that the high activity is of older date.

Other noteworthy centers of activity in the Indian Ocean are near the Gulf of Aden (Socotra), south of the Nicobar Islands (about 0°, 88°E.) and on the Kerguelen Rise.

The foregoing list of active belts accounts for by far the majority of recorded earthquakes, especially the larger ones. However, there are isolated areas, not evidently connected with any of the major active belts, in which either small shocks are frequent, or large shocks occur infrequently, or both. The last three centers of activity mentioned for the Indian Ocean are probably of this character. Such isolated areas exist in the Pacific; thus there is considerable small activity, and occasionally a moderately large shock, about the island of Hawaii. Scattered shocks occur in the region of Easter Island, and to some extent south and southeast of it in the area of the Easter Island Rise; epicenters in this region are usually not very accurate. Shocks also occur near the equator along the Galápagos Rise.

A good instance of a large shock occurring in a region where no previous activity was known is the Baffin Bay earthquake of Nov. 20, 1933.

The sporadic activity of eastern North America is not yet fully understood. Large shocks occurred in 1663 (Canada), 1811–1812 (central Mississippi Valley), 1886 (Charleston); and considerable shocks, for which instrumental epicenters are available, in southeastern Canada in 1925 and 1935.

The exact relationship of the great earthquakes of northwestern China (Kansu) and the Altai needs further discussion; they lie to the north of the Mediterranean or Alpide belt of structures.

As a general comment on the distribution of epicenters, it should be remarked that normal shocks, being associated with the most recent activity, do not always follow the older established tectonic lines, many of which originated in the Tertiary or earlier. Such lines are usually followed more closely by the epicenters of shocks at intermediate depths (see Chap. XI) and by volcanic vents. By a comparison of the recent with the older activity, the conclusion is reached that in the Pacific area the recent activity is displaced toward the Pacific basin. In the Alpide belt it is displaced toward the south.

As pointed out in Chap. X, the observed velocities of longitudinal and transverse waves are capable of yielding important information as to the crustal structure. For the few uppermost kilometers, which include the sedimentary layer where it exists, data are rapidly accumulating as a result of seismic field work using artificial explosions. In this way, local structures, faults, overthrusts,[19] have been investigated. Thicknesses of sediments of as much as 15 km. have been determined, so that it is often possible to follow the contour of the underlying Basement Complex. This level in continental areas either coincides with or immediately overlies the granitic layer of seismology.

Attempts to reach the base of this layer by the same methods have thus far been unsuccessful, and our seismic evidence as to the thickness of the granitic layer depends on the recording of natural earthquakes.

Seismic field work shows that, as was to be expected, the velocity increases with depth in any single material. However, in such work and in seismographic recording of large quarry blasts, it is often found that the first wave travels with a higher velocity than that found for longitudinal earthquake waves travelling in the granitic layer. It appears that some sedimentary rocks near the surface, especially those that are geologically old and consolidated, may transmit waves with higher velocity than granite.

Seismic methods have been applied by Ewing and his collaborators[6] in the region of the Atlantic coast of the United States, the experiments being continued out under water to about 100 km. from shore. The crystalline Basement Complex is found to be dipping gradually from a depth comparatively near the surface under the Piedmont Plateau (about 100 km. west of the shore line), to about 1,500 m. at the coast and about 4,000 m. at 100 km. off the coast. If these results are confirmed, they will prove highly significant.

The principles of the seismic method of investigation have two important further applications: in the first, the upper layer consists of water; in the second, it consists of ice. The very fruitful procedure of echo sounding is rapidly increasing our knowledge of the contours of the ocean bottom, which often has a bearing on speculation as to the underlying structures. Among the results are the contours of a great number of submarine canyons on the continental shelves (see Chap. III). Seismic methods have been used to determine the thickness of ice in glaciers,[21] in the vicinity of Little America in the Antarctic[22] and on the Greenland icecap.[23] For Greenland, the first discussion of the observations indicated the important conclusion that the base of the ice in the interior of the icecap is near sea level. This would suggest an isostatic depression of the rocks under the load of ice, analogous to that which is believed to have occurred in America and Scandinavia during the Pleistocene glaciations. However, the data have been rediscussed by Brockamp,[24] who concludes that no exact results can be derived from them for central Greenland, so that further observations will be required to determine the amount of the subsidence.

As already suggested, our principal source of detailed information on the structure of the crust at depths exceeding a few kilometers is the interpretation of seismograms of natural earthquakes. The velocities and structures derived from the study of the travel times of

earthquakes at the shorter distances are discussed in Chap. X. The observed amplitudes are capable of yielding further information. It can be shown theoretically[25] that a gradual increase of velocity with depth should be associated with large amplitudes at distances where the emerging ray reaches its deepest point at the depth where the increase in velocity occurs. If the velocity at the deepest point is nearly constant with depth, the corresponding emergent amplitudes are relatively small; if it decreases slightly with depth, the amplitudes are very small. Finally, if the velocity decreases with depth at a rate such that $dv/dr > v/r$ (where v is the velocity, and r is the radius vector from the center of the earth), the travel-time curve is interrupted, and a shadow zone exists in which only diffracted waves should be observed. The effect is of the same type as that of a discontinuity where the velocity is smaller in the lower medium; but the actual conditions require only a gradual decrease in velocity. Specifically, in the outer part of the mantle the velocity of longitudinal waves is between 8 and 10 km. per second, so that a decrease of only 0.02 km. per second, for an increase in depth of 10 km. is sufficient to produce a shadow zone.

This method of investigation has not yet been applied to waves observed at very short distances and consequently traveling within the continental crust. However, for waves that have penetrated to somewhat greater depths, data are available from various continental areas, particularly from California, Europe, and Japan. In all these regions the travel-time curves for the normal P (P_n) are practically rectilinear in the arc distance up to about 14°; the amplitudes are very small between 6 and 14°, apparently decreasing with distance in this range. At about 14° there is a sudden increase in the amplitudes, which continue large to distances well beyond 20°. From this behavior the conclusion appears unavoidable that at a depth of 50 or 60 km. the velocity ceases to increase with depth and that at slightly greater depth it even decreases with increasing depth. As the travel times show no perceptible delay at the distance of 15° where the amplitudes are large again, the total decrease in velocity cannot be large. As pointed out before, a very gradual decrease through a small range of depth is sufficient to produce the phenomenon of a shadow zone. The observations do not permit us to decide whether this decrease is gradual or sudden. The depth where the decrease occurs cannot be calculated with precision, as the numerical result will be much affected by slight inaccuracies in the observed times. In all probability the decrease begins at a depth in excess of 50 km., but not so much as 100 km.; this decrease cannot persist for more than 100 km., after which the velocity again begins to increase at a depth that is not much, if at all, in excess of 150 km.

As has been pointed out in Chap. VII, this phenomenon of decrease in velocity over a limited range of depth may well be due to a change of state, such as that from the crystalline to the vitreous condition.

At Pasadena it is a very frequent observation that longitudinal waves reflected at the surface of the earth (*PP*, etc.) are recorded with

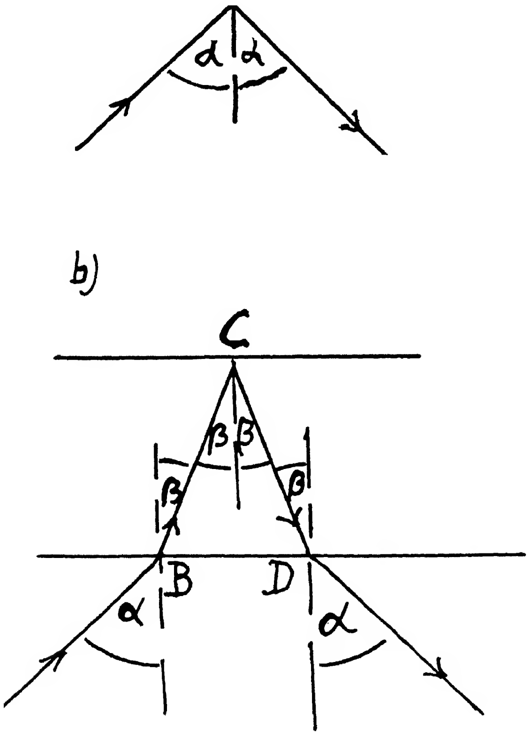

FIG. 23.—Angle of incidence for a certain distance. *a*) Wave reflected at the surface of the earth under the Pacific Ocean; *b*) wave reflected in a continent.

notably smaller amplitudes than the corresponding waves recorded in Europe from epicenters at similar distances. In the course of a systematic investigation it was found that this is always the case when the point of reflection falls in the Pacific basin. The theoretical explanation given by the present writers[25] is as follows.

The percentage of energy reflected when elastic waves are incident at a discontinuity depends principally on the angle of incidence. For an epicenter at a given distance, the angle of incidence at the surface of the earth is less in the presence of continental structure than when the continental crust is absent. These two cases are exhibited in Fig. 23, where *a*, represents the Pacific case and *b*, the continental case. The effect of structure on travel times and paths up to the first discontinuity is negligible, so that the angle α is the same in both figures if the epicenter is at the same distance. Since $\sin \beta/\sin \alpha$ is equal to the ratio of the two velocities, β is considerably smaller than α. In the case of teleseisms, the energy reflected at the surface of the earth increases greatly with decreasing angle of incidence. Calculations, which also took into account the loss of energy due to refraction at the base of the crust (*B* and *D* in the figure), showed that *PP* from a Pacific reflection should always be smaller than *PP* from a continental reflection at the same distance. The effect is most marked at distances between 40 and 50°, where the Pacific reflection on the average should have only about one-fourth the amplitude of a corresponding continental reflection.

Observation soon confirmed the occurrence of large amplitudes in *PP* at Pasadena in the comparatively few instances when the reflection is continental. Seismograms from other stations in all parts of the world were then examined, with the following general results.

Reflections of the Pacific type (small amplitude of *PP* or *PPP*) were found chiefly in the Pacific basin in the restricted sense. However, such reflections were also found in a limited area that coincides with the deepest part of the Arctic basin and in the deeper part of the southeastern Pacific off the South American coast. Continental reflections were found in the interior of all the continents, the continental seas of the Australasian area and without exception in the Atlantic and Indian oceans. Seismograms at Huancayo (Peru) exhibit continental reflections in the southeastern Pacific at about 25°S. 85°W., 20°S. 93°W. and 9°S. 91°W. The continental structure thus suggested is consistent with the numerous islands and submarine ridges existing in this part of the Pacific. All reflections from the ocean floor north and northeast of the Galápagos Islands are clearly Pacific.

In exceptional cases it is possible to locate the boundary between the two structures by this method. Thus seismograms at Pasadena show Pacific reflections taking place about 15°N. 98°W. and continental reflections about 15°N. 96½°W. Shocks from the Aleutian Islands and western Alaska regularly give Pacific reflections; but shocks farther

east show continental structure at the points of reflection, which then lie between 47 and 49°N., 129 and 134°W.

This method does not determine the thickness of the continental layer, which may possibly differ considerably in the various localities where the continental reflections are obtained. However, in every case the thickness must at least be a large fraction of the wave length of the longitudinal seismic wave, which in most cases is between 20 and 30 km.

The method of amplitudes of reflected longitudinal waves gives us information as to the structure at or near a single point in each case, *viz.*, the point at which the reflection occurs. Observations of surface waves, *i.e.*, waves that are propagated along the surface of the earth, can be made to yield similar information, which then refers to the integrated or mean condition along the whole path traversed.

In a medium that is not homogeneous, the velocity of surface waves depends on the period. Short waves are propagated only in a thin layer, the properties of which determine their velocity, whereas the energy of long waves is propagated in a relatively thick layer. In each case the disturbance extends from the surface to a depth that is a small multiple of the wave length; below this level the energy, being propagated horizontally, continues to fall off exponentially with depth and soon becomes negligible. The velocity of propagation is of the order of that of transverse waves; in general it is not far from the weighted mean of this velocity taken through the thickness of the disturbed layer.* If, for example, we have two layers; the upper one with a thickness of 10 km. and a velocity of 3 km. per second for transverse waves, the lower having a velocity of 4 km. per second, then surface shear waves with a period of 1 sec. will be propagated with a velocity of 3 km. per second, the wave length being of the order of 3 km.; if the wave has a period of 10 sec., the wave length will be greater than the thickness of the layer, so that a noticeable part of the energy will be propagated in the deeper layer and the velocity will be between 3 and 4 km. per second. Finally, if we consider a wave with a period of 60 sec., the wave length (nearly 240 km.) will be large compared with the thickness of the layer, nearly all the energy will be propagated in the deeper layer and the velocity will be nearly 4 km. per second.

If instead of two layers there are a larger number or if the velocity increases with depth, the effect will be similar, and the velocity of the waves will increase with period, as before.

* For the theory refer to Chap. 13, vol. VI, of this series, "Physics of the Earth—Seismology."

These general remarks apply to both of the two principal types of surface waves—the Love, or Q, waves, and the Rayleigh, or R, waves. The velocity of the latter is about 0.9 that of the former.

Figure 24 shows the variation of velocity with period for surface waves of both types which have traversed the different structural units named. This figure is reproduced from a previous publication by the present writers.[26]

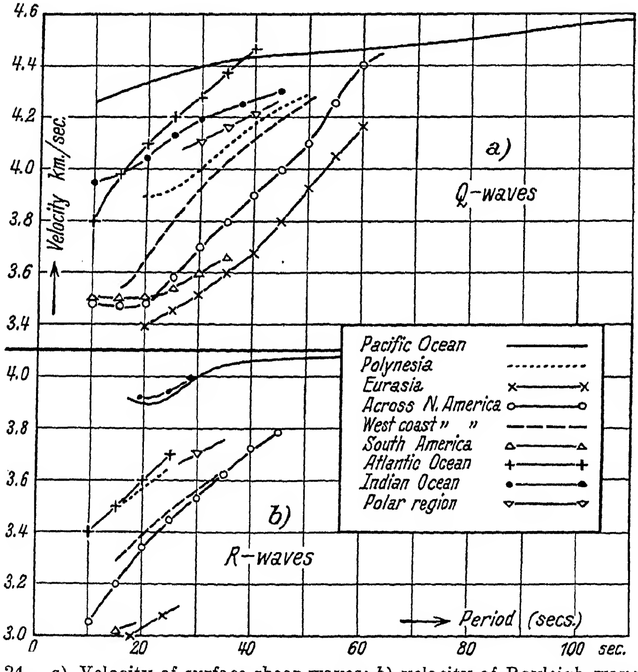

FIG. 24.—*a*) Velocity of surface shear waves; *b*) velocity of Rayleigh waves, along the surface of the earth in various regions as a function of their period. (*After Gutenberg and Richter.*)

The velocities in the Pacific basin are mostly higher than the others and show notably less variation with period, the slight increase with period in this case being fully capable of explanation as produced by an increase in velocity of bodily waves with increasing depth, due to the increasing pressure. These observations can be accounted for with the greatest degree of probability on the assumption[27] that the crustal layers characteristic of the continents are absent from the Pacific basin. However, the data are not inconsistent with an alternative explanation such as that suggested by Byerly,[28] who assumes a

surface layer in the Pacific area with a velocity for transverse waves of 4 km. per second, which is not very different from the corresponding velocity in the deepest of the continental layers.

In Fig. 24 the curves differing most widely from those for the Pacific paths are the curves referring to paths over the continents. Theoretically, the velocity of Q waves should approach 3.2 km. per second for short periods, since this is the velocity of transverse waves in the upper continental layers (excluding the sedimentary layer). The corresponding velocity for R waves should be about 3.0 km. per second. For very long wave lengths and periods the curves approach those for the Pacific, indicating a general uniformity at large depths. The fact that the Q curve for the west coast of North America is noticeably higher than the others suggests that the total thickness of the continental layers is thinner in these regions than in the others.

The remaining curves represent paths which are wholly or in large part oceanic but which lie outside the Pacific basin; they lie between the continental groups of curves and those for the Pacific. This may mean either that the total thickness of the layers of continental type in these oceanic areas is thinner than on the continents proper or that the upper layers, with their lower velocities, are missing, leaving an intermediate continental layer at the surface. The fact that the velocities for short periods are higher favors the latter explanation. We are inclined to apply this to the Atlantic and Indian oceans, and possibly to the Polynesian area. It has been mentioned that in the north polar region the amplitudes of reflected longitudinal waves indicate the existence of a limited area in which the structure is of Pacific type. The data from surface waves do not contradict this, as they refer to paths only a fraction of which lie through the limited area in question, the remaining segments being through unquestionable continental regions. The velocities through the polar basin itself must be higher than those given by the data of Fig. 24 and probably are close to those in the Pacific basin.

Theoretically it should be possible to draw conclusions regarding the crustal structures from the ratio of the horizontal to the vertical component in the observed Rayleigh waves and also from the distribution of periods in surface waves of all types (including the cauda or tail waves of the seismograms) in various regions. For both these purposes the available data cannot yet be considered sufficient.[26]

There is a considerable loss of energy when short surface waves cross the boundary of the Pacific basin. Neither the boundaries of the Atlantic nor those of the Indian Ocean show a similar effect.[26,29] Although this effect is strong for waves with periods of 20 sec. (wave

lengths about 70 km.), it decreases with increasing period, so that waves with periods about 1 min. (lengths of about 250 km.) show no appreciable loss of energy in crossing the boundary of the Pacific basin. This indicates that the structural discontinuity does not extend to depths of the order of 200 km.

CONCLUSIONS

All evidence agrees in dividing the surface of the earth into two areas which are characterized by different structures. The first of these includes the Pacific basin, possibly with a few outlying regions, one of which is in the Arctic basin. The second comprises the remainder of the surface, with the present continents and continental seas, the Atlantic and Indian oceans and possibly isolated patches within the borders of the Pacific area (see Fig. 20).

The margins of the Pacific basin are largely parallel to mountain chains which are frequently folded in arcs convex toward the basin. Such structures are conspicuously absent from the coasts of the Atlantic Ocean and most of the Indian Ocean. However, they occur along the boundaries of the Mediterranean or Alpide belt; this includes the arc of the Malay Peninsula, Sumatra, etc., limiting the Indian Ocean on the northeast and continuing in a traceable curve northward through the East Indies and by way of the Philippines to Japan, where it joins the Pacific belt. The area of the western Pacific Ocean lying between these two belts (between the Marianas Islands and the Philippines) is part of the area of continental structure type and is not to be included in the Pacific area proper. The western and southwestern boundary of the Pacific area is indicated by the andesite line; outside this boundary the younger eruptive rocks are chiefly andesitic, whereas inside the Pacific area they are basaltic.

The margin of the Pacific area is associated with great seismic activity. The actual boundary is most nearly outlined by the epicenters of the larger normal shocks; but such shocks occur in other regions, particularly in the Mediterranean-Alpide belt. Shocks at great depth (300 to 700 km.) are known only from the circum-Pacific belt, where their epicenters always lie outside the Pacific area within the surrounding continental structures. Their alignment is not always associated with the present boundary of the Pacific area; thus the belt of deep shocks paralleling the line of the Marianas and Bonin Islands continues directly across central Japan and the Japan Sea into Manchuria, and the very deep shocks occurring in the Philippines and south of Borneo are aligned parallel to the Alpide structures.

It appears that the boundary between the two chief structural units, Pacific and continental, extends to depths of at least 40 km., and that in some way it affects conditions down to at least 700 km. Seismic data show no significant horizontal difference in physical properties between the continental and Pacific areas, below a depth of about 50 km. However, the difference in structure near the surface is sufficient to cause a large absorption of the shorter surface waves crossing the boundary.

The distinction between the two chief structures is supported by determinations of the mean velocity of surface waves over paths of different character (Pacific or continental), by a few observed velocities of bodily waves in the vicinity of Samoa (see Chap. X) and by readings of the relative amplitudes of reflected longitudinal waves (*PP*, etc.). The results consistently indicate that in the Pacific area the material near the surface does not differ significantly in physical proportions from that several hundred kilometers below it, a condition quite different from that in the continental area. The rigidity and bulk modulus are higher for Pacific rocks than for continental rocks (Chap. XIV).

The typical continental structure consists of a series of layers, the uppermost of which is that of the sedimentary rocks (including metamorphosed sediments). The elastic constants within this layer, as well as its thickness, are subject to much local variation. Over large areas the sediments are completely absent, whereas in deep basins of geosynclinal character they may extend to depths of as much as 15 km. Where the sediments are absent, the exposed rocks are usually granitic (basaltic in volcanic regions); moreover, in many localities granitic rocks are known to underlie the existing sediments. The velocity of longitudinal waves in unweathered granite near the surface is about $5\frac{1}{4}$ km. per second. Seismic data give a velocity of about 5.5 or 5.6 km. per second for the entire layer of which the surface granites appear to be a part. The difference is attributable to increased pressure at depth, so that it is usual to refer to the layer in which this velocity occurs as the granitic layer.

In the continental regions where it has been studied, the base of the granitic layer has been found at depths between 10 and 30 km. Below this is a small number of layers (perhaps only one in some regions). The thicknesses and elastic properties apparently differ notably in the various regions. It has been suggested that in this intermediate layer or layers the material is basaltic; but this conclusion has been questioned by several authors. The lower boundary of these layers is the Mohorovičić discontinuity (Chap. X). This is a major discontinuity,

which has been found so far in all continental regions where data are available. Its depth, which is the total thickness of the continental structures, has been determined at from 30 to 50 km. This thickness seems to be relatively small in California, western Europe, New Zealand and northeastern Japan; about average thicknesses occur in central North America and South America. The largest values found thus far are in the region of the Sierra Nevada, California (40 to 65 km.[39]), the Alps (50 km.) and, probably, the Caucasus. These differences within the continental area are related to differences in geological history. Certain regions are or have been covered by continental seas; others have been disturbed by mountain building, although the mountains may have been removed by later erosion. The greatest outstanding zone of disturbance is the Mediterranean, or Alpide, belt of Tertiary mountain building, which is accompanied by volcanism and seismic activity, including intermediate shocks as well as shocks at normal depth.

Within the general area of continental structure are several oceanic basins which differ from the continental seas; they are the Atlantic and Indian oceans and a large part of the western Pacific Ocean. Data of all kinds are most abundant for the Atlantic Ocean.

In general, the coasts of the Atlantic Ocean are not paralleled by lines of tectonic, seismic or volcanic activity. Structurally, the boundary of this ocean basin is not a sharp discontinuity like that delimiting the Pacific basin but appears to be a gradual transition. Surface waves are transmitted across the boundary with no appreciable loss of energy, indicating that there can be no important discontinuity. On the North American coast the structures have been followed by geophysical and geological methods and appear to dip under the ocean bottom without any break.

There are exceptional circumstances, such as the Antillean arc and the analogous arc extending into the south Atlantic from Cape Horn, in which the boundary of the Atlantic basin is followed by tectonic structures; but these form only a small fraction of the whole.

The Mid-Atlantic Ridge is outlined by numerous epicenters, especially from the equatorial region northward; it is probably a line of present tectonic activity.

Reflections of longitudinal waves indicate the presence of continental layers in the Atlantic; the study of surface waves indicates that the total thickness of these layers is less than on the continents and that possibly the uppermost (granitic) layer is very thin or absent.

The geological bearing of these results has been discussed by Field,[30] who summarizes his conclusions as follows:

The analysis of earthquake waves, therefore, not only suggests the Atlantic to be everywhere underlain by a considerable thickness of continental rocks, but also intimates that the Pacific continental margins are of much greater structural significance than those of the Atlantic. This suggestion is still further strengthened by the recent "artificial seismic" profile from the edge of the Atlantic Cretaceous overlap to the continental margin, where the sedimentary rocks composing the bathyal slope have been proved to be only one-half the thickness of the stratigraphic column overlying the greatly downwarped pre-Cretaceous contact. . . . It would appear as if the present topographic limit of the Atlantic continental margin had no particular structural significance; that is, so far as ancient structural delimitation of continental block and ocean basin is concerned. It is possible, therefore, that the sub-Atlantic lithosphere constitutes a vast area of *downwarped* pre-Cambrian and Palaeozoic geology, fully comparable in the complexity of its subsidiary stratigraphic, structural, and palaeogeographic features to the *upwarped* pre-Cambrian and Palaeozoic geology of the surrounding continental areas. If this be true, Schuchert's necessary but modest postulate of a limited Appalachia is but a structural element of the whole of the sub-Atlantic lithosphere, and stratigraphers are finally relieved of the insupportable structural difficulties which they now encounter in their Palaeozoic palaeogeographic problems. It is not implied that the downwarp of this great pre-Mesozoic continent of Atlantica was relatively rapid, or that it took place all at once. In the late Tertiary there may still have been some sub-aerial remnants whose land areas were temporarily enlarged by the particular climatic events of the Pleistocene. What *is* implied, however, is that, through the pre-Cambrian and Palaeozoic history of the earth, the Atlantic region was characterized by seas, lakes, rivers, mountains, sediments, marine and terrestrial organisms respectively similar to, and coexistent with, those of the continents.

The geographical and geological evidence for the Indian Ocean is similar to that for the Atlantic; the coasts are of Atlantic type, except in the northeast sector, where they follow the structural line of the Mediterranean-Alpide orogeny, which extends along the Malay Peninsula, Sumatra, Java, etc. The seismic and geophysical data are scanty but where available are of the same character as for the Atlantic. Earthquake epicenters along the boundaries of the ocean basin are rare, except along the structural line just mentioned, which is also accompanied by deeps and negative gravity anomalies; its position is analogous to that of the West Indian arc in the Atlantic area. This line continues into the tectonically disturbed East Indian region, where the conditions reach an extreme of complexity about the Banda Sea. Everywhere the seismic data indicate continental structures, which probably vary greatly in detail from point to point. In this area all types of earthquakes occur—at normal, intermediate and great depths—as well as volcanism and large gravity anomalies.

The tectonic line that has been mentioned is closely paralleled by the belt of negative gravity anomalies discovered by Vening Meinesz. This continues round the Banda Sea and then curves northward to the east coast of the Philippines; the structures, whether associated with gravity anomalies or not, are continuous northward to the coast of Japan, where they reach the boundary of the Pacific area. The western portion of the Pacific Ocean, lying between the line passing east of the Philippines and that part of the andesite line which extends through the Bonin and Marianas Islands, is shown by the seismic data to be an area of continental structure. The data are not sufficient to show whether the structures in this region are analogous to those of the Atlantic or whether they correspond to those of the land areas.

Gravity observations are consistent with the general description of structures thus far outlined. Where the requirements of isostasy are fulfilled (which is usually the case), the observed gravity is consistent with continental structure as described above. Where there are large departures from isostasy (large gravity anomalies), there is usually high seismicity and other evidence of recent tectonic activity. The data indicate that the plastic flow required to maintain isostatic equilibrium may occur at depths as shallow as 50 km. (see Chap. XV).

At depths between 60 and 80 km. the velocity of longitudinal and of transverse waves seems to decrease slightly with increasing depth, perhaps as a consequence of a transition of crystalline to vitreous material.

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CHAPTER XIII

DENSITY, GRAVITY, PRESSURE AND ELLIPTICITY IN THE INTERIOR OF THE EARTH

WALTER D. LAMBERT

(Senior Mathematician, U. S. Coast and Geodetic Survey)

If we assume hydrostatic equilibrium within the earth and a law giving the density at any distance from its center, then the equipotential surfaces within the earth are also surfaces of equal pressure and equal density, and this hydrostatic pressure is determinate. Under the same conditions the acceleration of gravity is also known. But if we abandon the hypothesis of equilibrium, the level surfaces are no longer surfaces of equal density and the idea of a surface of equal pressure becomes meaningless, for we would then be dealing with all the complicated phenomena of stresses in an elastic or elasticoviscous solid. The values of gravity within the earth would, however, be dependent only on distribution of density and would be unaffected by the existence or nonexistence of hydrostatic equilibrium.

The assumption of hydrostatic equilibrium within the earth is probably not far from the truth at moderate depths, in spite of the existence of deep-focus earthquakes. In what follows, an attempt will be made to distinguish between the conclusions based on the assumption of hydrostatic equilibrium and those conclusions for which this assumption is not needed.

THE MEAN DENSITY OF THE EARTH

The mean density ρ_m and the Newtonian gravitational constant k are not known independently, but the product $k\rho_m$ is known to a high degree of accuracy. If we denote by R the mean radius of the earth and neglect its rotation and ellipticity and if we call G the acceleration of gravity on its surface, we have by the law of the inverse square for a sphere

$$\frac{4}{3}\,[illegible]\,\frac{[illegible]}{R^2}$$

or

$$[illegible] = \frac{[illegible]}{4\pi R}. \qquad (43)$$

To take account of the flattening and of the rotation, let g_e be gravity at the equator, e the flattening and a the equatorial radius; we then have

$$\frac{3g_e}{4\pi a}\left[1+\frac{3}{2}m\left(1+\frac{2}{7}e+\frac{125}{441}e^2\right)\right]. \tag{44}$$

Here m is the ratio of the centrifugal acceleration at the equator to gravity at the equator, or $m = \omega^2/g_e$, where ω is the angular velocity of rotation.*

The formula out to only the term $\frac{3}{2}m$ is amply accurate for most practical purposes in view of the accuracy of the data. There is then no need to know the exact flattening. If we include the term in e, we still do not need to know within 1 part in a million whether the surface is an exact ellipsoid or not.

The Newtonian gravitation constant is determined independently by the Cavendish experiment. The most recent determination is by Heyl.[1] His result is 6.670×10^{-8} c.g.s. unit.

The value of gravity at the equator is taken as 978.049 gals in the international gravity formula, and this is as good as any, provided that we accept the determination of absolute gravity at Potsdam on which it is based and which has been for many years accepted as the best available. Heyl and Cook, however, made a determination of absolute gravity at Washington,[2] and, on the basis of this and of several good gravity connections between Washington and Potsdam, Heyl and Cook concluded that the value of absolute gravity at Potsdam and hence the value of g_e in the international formula should be reduced about 0.020 gal. The correction determined by Heyl and Cook has been substantially confirmed by recent work at the National Physical Laboratory, Metrology Department, at Teddington, England, which obtained a correction of 0.013 gal[2a], corresponding to $g_e = 978.036$. It should be remembered that most geophysical inferences are based on the relative values of gravity or on the gravity anomalies and that neither would be altered by a change in the absolute value of gravity.†

The data are so uncertain and the mean departure of the earth from an exact ellipsoid so slight that we shall use the formula

(45)

* The formula assumes that the earth is an exact ellipsoid.

† The determination of absolute gravity is a long, difficult and tedious operation, and Heyl and Cook do not claim an accuracy that would attribute much meaning to the third decimal of a gal.

with the values of g_e both from the international formula as it stands and is corrected in accordance with the work of Heyl and Cook and with value of a from the international ellipsoid of reference adopted by the International Association of Geodesy at its Madrid meeting in 1924, *viz.*,

$$a = 6{,}378{,}388 \text{ m.} = 6.378388 \times 10^8 \text{ cm.}$$

We find that for the accuracy commonly needed there is very little difference, owing to the different values of g_e. We have for $g_e = 978.029$ gals (international gravity formula with absolute gravity corrected by Heyl and Cook)

$$\begin{aligned} \log k\rho_m &= 3.5658057 - 10. \\ \rho_m &= 5.5167. \end{aligned} \tag{46}$$

For $g_e = 978.049$ gals (international gravity formula based on absolute gravity determined at Potsdam by Kühnen and Furtwängler),

$$\begin{aligned} \log k\rho_m &= 3.5658145 - 10. \\ \rho_m &= 5.5168 \end{aligned} \tag{47}$$

It is seldom that we need to discuss the mean density of the earth beyond its fourth decimal.

GRAVITY AND PRESSURE IN THE INTERIOR OF THE EARTH

Gravity in the interior of the earth does not depend in any great degree on the fluidity or nonfluidity of the interior of the earth. Apart from the ellipticity, it depends on the law of variation of density from center to surface. Let ρ be the density at distance r from the center. The mass of the sphere of radius b is then

$$r^2\, dr, \tag{48}$$

and gravity g at distance b is given by

$$g = \frac{4\pi k}{b^2} \int_0^b \rho r^2\, dr, \tag{49}$$

because the spherical shell (we are neglecting the flattening) between the radius b and the outermost radius a exerts no attraction on a point within the shell.

From Eq. (49) it follows that g has a maximum at a depth d, if the density there equals $\frac{2}{3}$ of the average density of the part of the earth below the depth d. As, near the surface, the density is smaller than the critical value ($\frac{2}{3} \times 5.52$), gravity must increase with depth

in the surface layers. Its maximum is probably at the surface of the core.

It is usual to assume hydrostatic equilibrium within the earth. To trace the consequences of this assumption let p be the pressure at distance r from the center. From the well-known relation

$$dp = -g\rho\, dr \tag{50}$$

we find for the pressure at distance b from the center

$$-p = 4\pi k \int_r^a \rho \frac{dr}{r^2} \int_0^r \rho x^2\, dx, \tag{51}$$

where a is the outer radius of the earth and r as a variable of integration has been replaced in the right-hand integral by x. We can obtain a differential equation simple in appearance by introducing the new variable $\bar{\omega}$ connected with p by the relation

$$dp = \rho\, d\bar{\omega} \tag{52}$$

and differentiating twice. After some transformations we find

$$\frac{d^2}{dr^2}(\bar{\omega}r) + 4\pi k\rho r = 0. \tag{53}$$

THE ELLIPTICITY INSIDE THE EARTH

On the hypothesis of hydrostatic equilibrium throughout the interior, the ellipticity is connected with the distribution of density by the celebrated differential equation due to Clairaut.[3,4,5] To abbreviate, let ρ be the density at distance r from the center, r being in strictness the mean radius of the ellipsoid of ellipticity e, and let M for brevity be given by

$$M = \int_0^r \rho x^2\, dx. \tag{54}$$

It is seen that apart from the constant factor 4π the quantity M denotes the mass within a sphere of radius r. Then Clairaut's differential equation is

$$\frac{r}{M} \qquad\qquad - \frac{6}{2}\Big) e =$$

This equation is accurate to the small quantities of the order e only. It makes no distinction between spheroids of revolution that are exact

ellipsoids of revolution and those that are not. The quantity e is defined by

$$e = \frac{a - b}{a}, \tag{56}$$

where a and b are the semiaxes of the spheroid within the earth and where in reality we are making no very rigid distinction between $(a - b)/a$ and $(a - b)/b$, so that the reciprocal of e is uncertain by about a unit. In spite of these simplifications, the solution of Clairaut's equation in an easily interpretable form has been obtained in only a few cases. The best known solution is that of Legendre for the law of density.

$$\rho = \frac{\rho_0 \sin nr}{nr}, \tag{57}$$

where ρ_0 is the central density. Before proceeding to show that this equation gives an expression for the ellipticity in finite terms, we shall state some of its consequences. We find

$$4\pi k\rho_0\left(\frac{nr - nr \cos nr}{r^2}\right). \tag{58}$$

$$n\,(\rho^2 - \rho_1^2) = \frac{2\pi k\rho_0^2}{n^3}\left(\frac{\sin^2 nr}{r^2} - \frac{\sin^2 na}{a^2}\right), \tag{59}$$

where ρ_1 is the surface density and a the mean radius of the surface $= 6371.2$ km. The result for the flattening is more complicated. The flattening of any surface with mean radius r is proportional to

$$\frac{(3 - n^2r^2) \tan nr - 3nr}{n^2r^2(\tan nr - nr)}.$$

The factor of proportionality is somewhat complicated and involves the quantity m, the ratio of the centrifugal acceleration at the equator to gravity at the equator, also the quantity $na = \nu$, which is the value of nr at the surface.* The resulting ellipticity (flattening) e, for any radius r, is

$$e = \frac{5}{2}m\frac{(\tan \nu - \nu)^2}{(\nu^2 - 2) \tan^2 \nu + \nu \tan \nu + \nu^2} \cdot \frac{(3 - n^2r^2) \tan nr - 3nr}{n^2r^2(\tan nr - nr)}. \tag{60}$$

* The transformations are long and tedious and require some mathematical ingenuity.

The value of $na = \nu$ that gives a flattening of $\frac{1}{297}$ is

$$\nu = 2.516503 \text{ radians} = 144° \ 11' \ 06''.$$

This corresponds to a central density of 11.2 and a surface density of 2.6, which seems somewhat too low.

If we assume the seemingly complicated law

$$\rho = \rho_0(1 - cr^\lambda)^{\mu-1}\left[1 - c\left(1 + \frac{\lambda\mu}{3}\right)r^\lambda\right] \tag{61}$$

where ρ_0 is the central density and c, λ and μ are three disposable parameters, it is found that the ellipticity e can be expressed as a hypergeometric series in powers of cr^λ. This has the advantage that the properties of the hypergeometric function are well known and that there are more disposable parameters to represent conditions within the earth than is the case with Legendre's law. Nevertheless the law of density, which in most general form is due to Maurice Lévy,[6] has not been much used in its most general form but is chiefly known in its very special form of Roche's law.

$$\rho = \rho_0(1 - c_1r^2), \tag{62}$$

c_1 being a constant for $r < a$. For a discussion of the mathematical analysis of Roche, Lipschitz and Lévy and for references to the original sources see Tisserand, "Mécanique céleste," vol. II, Chap. XV.

THE DENSITY INSIDE THE EARTH

The laws of density so far mentioned are now of interest chiefly for mathematical or historical reasons. They were developed at a time before seismology had given us some insight into conditions within the earth. Seismological evidence shows that there are discontinuities in physical properties at one or more depths within the earth, whereas the foregoing laws assume continuity, as if of a single substance.

Two papers of interest, by Wiechert[7] and Klussmann,[8] consider what was then known from seismological evidence regarding the interior of the earth. It should be remembered also: (1) that the mathematical difficulties are great and that the simplifying assumptions were made from necessity rather than from choice; (2) that one of the principal purposes of both papers was to compute the deviation of the outer surface from an exact ellipsoid, on the hypothesis of internal hydrostatic equilibrium, a question of no particular interest to us in the present connection. With these points in mind, it is of interest to examine some of the conclusions reached.

Wiechert presents the evidence in favor of a nucleus of essentially different construction from the outer layers and suggests that this nucleus might be iron or nickel or a combination. For mathematical simplicity he confines himself to the case of a nucleus and an outer shell, each of constant density throughout. This density represents, of course, some kind of mean density. He makes three assumptions as to the density of the shell. From his Tables II*a* and II*b* (Ref. 7, page 230) the following figures are taken for surface ellipticity = $\frac{1}{297}$.

Density of outer shell ρ_1	Density of nucleus ρ_2	Equatorial radius of nucleus r_2, km.	Reciprocal of ellipticity of nucleus e_2	Precession constant $(C - A)/C$
3.0	8.046	5,100	320	305.204
3.2	8.206	4,978	323	305.203
3.4	8.423	4,829	328	305.200

The quantities C and A are principal moments of inertia of the earth; their importance will appear later.

Klussman considers a three-layer earth, consisting of a nucleus, an intermediate shell and an outer shell. He takes fixed equatorial radii 3,924, 5,185 and 6,378 km. for the outer equatorial radii of the nucleus and the shells. He assumes a flattening of $\frac{1}{298}$ instead of Wiechert's $\frac{1}{297}$ and a mean density of 5.53, as against Wiechert's 5.56, but these underlying data are nearly enough equal to render the unrevised results fairly comparable.

Density of outer shell ρ_1	Density of intermediate shell ρ_2	Reciprocal of ellipticity of intermediate shell $1/e_2$	Density of of nucleus ρ_0	Reciprocal of ellipticity nucleus $1/e_0$
3.0	7.224	322	8.340	327
3.2	6.606	324	8.750	335
3.4	5.987	327	9.161	343
3.6	5.366	330	9.575	353

Klussman does not give the precession constant $(C - A)/C$, but Radau's transformation, to be explained later, shows that this varies within a very narrow range in direct proportion to the ellipticity of the surface, almost regardless of the law of internal density, provided that

hydrostatic equilibrium be assumed. Wiechert's numerical computations for various ellipticities with various surface densities and Tisserand's computations with Legendre's very different law of density both illustrate this fact and show that $(C - A)/C$ varies about one and one-fourth times as fast as $1/e_1$, the surface ellipticity. Thus the value of $(C - A)/C$, which may be derived from astronomical data, practically determines the ellipticity of the surface, almost independently of the law of distribution of density, if we adopt the hypothesis of hydrostatic equilibrium. Wiechert considers the evidence then available and concludes that the hypothesis of hydrostatic equilibrium is compatible with it but that, on the other hand, after allowing for uncertainties in the data, the surfaces of equal density *may* deviate from the level surfaces within the earth by 100 m. or more. Today we incline to the same conclusion, with more emphasis on the probably close agreement between the flattening deduced geodetically and that deduced astronomically from $(C - A)/C$ and with less emphasis on a probable disagreement.

The question may legitimately be asked whether the precession of a fluid spheroid is the same as that of a solid one having the same moments of inertia. The answer is "yes, for all practical purposes." The question has been investigated by Darwin.[9] Schweydar[10] reaches a similar conclusion in regard to the precession and nutation of an elastically yielding earth. It appears, therefore, that the yielding of the earth, whether elastic or viscous, has little effect on the precession and nutation, neither has the distribution of density within the earth, if the earth be approximately in hydrostatic equilibrium. The surface ellipticity (e_1) and the precession factor $\left(\frac{C - A}{C}\right)$ are then very nearly determined one by the other, and any disagreement in the conclusions drawn from the observed values of these quantities, having regard to their probable errors, throws very little light on the distribution of density within the earth.

CLAIRAUT'S AND RADAU'S EQUATIONS

As a preparation for Radau's transformation of Clairaut's differential equation we introduce two formulas of some interest in themselves. They refer to the rate of increase of ellipticity with depth. Their chief application is near the surface. If we write the expression for gravity at the surface in latitude ϕ

$$g = g_e(1 + \beta \sin^2 \phi), \tag{63}$$

which is accurate to the order of small quantities here retained, we have for the quantity β

$$\beta = \tfrac{5}{2}m - e_1, \tag{64}$$

where m has been defined and e_1 is the ellipticity of the outermost surface. This is Clairaut's finite equation. From the consideration that, regardless of density, the work done in descending to a level surface close to the outermost surface varies for a given distance in proportion to superficial gravity, that consequently the distance between adjacent level surfaces in different latitudes is inversely as gravity and that therefore adjacent level surfaces at the pole and equator are not equidistant and have different ellipticities, we readily find for the rate of change of the ellipticity (Ref. 4, vol. II, page 93) for distance

$$de \quad 1$$

$$- 2e_1\Big). \tag{65}$$

Incidentally this equation, being independent of the density, holds good inside or outside the free surface. An equation for the second derivative may be mentioned in this connection, though not needed in the derivation of Radau's transformation. It is

$$a^2 \frac{d^2e}{dr^2} = 6\left(e_1 - \frac{\rho_1}{\rho_m}\beta\right), \tag{66}$$

where ρ_1 is the surface density (Ref. 4, vol. II, page 487). Other needed equations give relations between the moments of inertia and and the quantities e_1 and m. They are

$$\left.\begin{aligned} \frac{C - A}{} &= \frac{2}{3}\left(e_1 - \frac{1}{2}\right. \\ C &= \tfrac{8}{3}\pi \int_0^a \rho r^4\, dr. \\ M &= 4\pi \int_0^a \rho r^2\, dr. \end{aligned}\right\} \tag{67}$$

That is, the factor 4π has been introduced into the quantity previously[4] defined as M.

These give

$$\frac{C - A}{} = \frac{\left(e_1 - \frac{m}{2}\right)a^2 \int_0^a \rho r^2\, dr}{}.$$

To introduce Radau's[11] ingenious and enlightening transformation of Clairaut's differential equation, we introduce the mean density D of a spheroid of radius r. This mean density is, of course, the mass divided by the volume. The constant factor 4π drops out, and, neglecting the flattening because the mean radius is used, we have

$$Dr^3 = 3\int_0^r \rho x^2\, dx = \rho r^2 + \int_\rho^{\rho_0} x^3 \frac{d\rho}{dx}\, dx, \tag{69}$$

or by differentiating,

where ρ is the density at radius r; the second form is obtained by an easy integration by parts and x is a variable of integration, the limits of which are 0 and r corresponding to densities ρ_0 and ρ. By substitution in Clairaut's equation (55), we easily find

Radau further introduces a new variable η, defined by

$$\eta = \frac{r}{e}\frac{de}{dr}, \tag{72}$$

from which is found

$$\frac{de}{dr} = \frac{e}{r}\eta.$$

$$\frac{d^2e}{dr^2} = \frac{e^2}{r^2}\left(r\frac{d\eta}{dr} + \eta^2 - \eta\right). \tag{73}$$

Expressing the results in terms of D and η, we get

$$\frac{d\eta}{dr} + \eta^2 + 5\eta\Big)\, D + 2r(1 + \eta)\frac{dD}{dr} = 0. \tag{74}$$

This equation Radau writes in the form

$$- \eta^2 \qquad = 0. \tag{75}$$

We shall need $(d/dr)(r^5\sqrt{1 + \eta}\, D)$ for a reason that will appear. Treat the quantity to be differentiated as the product of r^5 and $\sqrt{1 + \eta}\, D$. We find

$$d \qquad\qquad\qquad\qquad d \qquad\qquad D$$

$$+ \eta^{(5\eta + \eta^2)\, D}$$

$$= 5r^4\, D\frac{1 + \frac{1}{2}\eta - \frac{1}{10}\eta^2}{} \tag{76}$$

The key to the usefulness of Radau's transformation lies in the fact that the fraction $(1 + \frac{1}{2}\eta - \frac{1}{10}\eta^2)/(\sqrt{1 + \eta})$ is very near unity for a wide range of values of η. By expanding in series its value to terms in η^3 inclusive is $1 + \frac{1}{40}\eta^2 - \frac{3}{40}\eta^3 \ldots$, a small quantity even for $\eta = 0.5$ or 0.6, which appears to be about the largest reasonably probable value for η. Numerical computation tells the same story. Integrate (76) from center to surface and let η_1 represent the surface value of η. The value of D for the surface is simply $D_1 = \rho_m$, the mean density.

We find

$$= 5\int_0^a r^4\, D\frac{1}{} \qquad dr. \tag{76a}$$

It can be shown that at the center of the earth $\eta = 0$ and for the surface by Eqs. (65) and (72)

$$\eta_1 = \frac{5}{2}\frac{m}{e} - 2 = 0.57, \tag{77}$$

so that as a close approximation we may consider that the fraction $(1 + \frac{1}{2}\eta - \frac{1}{10}\eta^2)/(\sqrt{1 + \eta})$ hardly affects the integrand and we write

$$5\int_0^a r^4 dr = a^5 \qquad \overline{\cdot\eta_1}, \text{ nearly} \tag{78}$$

We have

$$= \rho_m \tag{79}$$

Multiply by $d(r^2)$, and integrate by parts.

$$^4\, D\, dr = \int_0^a r^3\, D\, d(r^2) = a^5\, D_1 - 3\int_0^a \rho r^4\, dr. \tag{80}$$

By using Eqs. (68), (77), (79) and (80), also (78) with $r = a$, we find

where f is a small quantity representing the error introduced by equating $(1 + \frac{1}{2}\eta - \frac{1}{10}\eta^2)/(\sqrt{1 + \eta})$ to unity. The quantity f is 0.0004 or less.

This elaborate and ingenious transformation explains in a way why the surface ellipticity practically determines the precession regardless of the law of density, hydrostatic equilibrium being assumed, but it gives little or no physical insight into the problem.

The best statement of the physical reason is paraphrased from Heger[12] by Helmert (Ref. 4, Vol. II, page 491). We may paraphrase is thus: If in Eq. (68) we assume constant density with ellipsoidal layer of constant flattening, we get $e_1 = \frac{1}{270}$, approximately, from the known value of $(C - A)/C$, which is very nearly the actual flattening. If we know from Eq. (65) that the ellipticity of the inner level surfaces begins by decreasing with depth and it takes only a small decrease to reconcile the observed ellipticity and the observed precession constant, $(C - A)/C$.

The agreement of the ellipticity and the precession constant for any assumed law of density is not an argument in favor of that law. On the other hand, a flat disagreement in excess of what might be accounted for by errors of observation might indicate the absence of hydrostatic equilibrium within the earth.

NUMERICAL DATA

A numerical tabulation of the density, gravity and ellipticity within the earth according to various assumptions may be of interest. Three sets of figures are available, based on: (1) Wiechert's law with the mean density reduced from Wiechert's rather high value of 5.58 down to the more modern value 5.52; (2) Legendre's law with $\nu = na$ taken as 143° exactly, instead of 144.°185, as given in the preceding text, the figures for $\nu = 143°$ being taken because immediately available and sufficiently accurate for purposes of illustration and comparison; (3) figures from two papers by Bullen[13] in which use is made of seismological data and the figures are obtained by mechanical quadrature. Bullen's papers seem to present adequately the state of today, and accordingly a brief summary of them is given in Table 63.

The figures for the density and pressure are interpolated from Bullen's second paper, the argument being changed from depth to distance from the center in terms of the earth's radius as unity. In his first paper he gives the ellipticity itself, from which the reciprocal of the ellipticity as far down as the central core has been computed, since Bullen in his second paper states that his recomputation made no substantial change in the figures for the ellipticity for this part of the

TABLE 63

DENSITY, PRESSURE AND ELLIPTICITY WITHIN THE EARTH (BULLEN)[13] AND GRAVITY (BENFIELD[14])

Distance from center r/a	Density ρ gm./cm.^{3}	Pressure p, bars*	Reciprocal ellipticity $1/e$	Gravity, g, gals.
Center 0.0	12.2	3.5×10^6	400	0
0.1	12.1	3.4	398	214
0.2	11.9	3.2	396	426
0.3	11.4	2.7	394	623
0.4	10.8	2.2	392	809
0.5	10.1	1.63	390	973
0.6	5.4	1.15	379	1007
0.7	5.1	0.87	360	980
0.8	4.7	0.52	333	981
0.9	4.3	0.23×10^6	313	992
Surface 1.0	....	....	297	982

* 1 bar = 1 million dynes/sq. cm. = 0.987 atmosphere.

earth. Bullen does not give precise figures for the ellipticity of all strata in the core; the figures here set down are partly estimates based on his statements regarding his recomputation.

Incidentally Bullen has verified the fact that for the distribution of density here involved the approximation

$$1 + \tfrac{1}{2}\eta - \quad 1$$

is a close one.

The major discontinuity between core and mantle occurs between $r/a = 0.5$ and $r/a = 0.6$.

For comparison there are given the results of a computation similar to Bullen's but with Legendre's law of density. This law was long in favor because of its mathematical convenience and was, of course, used a century before there was any idea of a discontinuity between core and mantle. In spite of this fundamental difference the results of the two methods of computation present no striking contrast. Legendre's law gives the surface density as too low according to modern views and the density at depths corresponding to the lower parts of the mantle rather too high.

Table 64 gives a possible distribution of density, pressure, etc., closely satisfying all known data except those of seismology. The mathematical formulas on which it is based are exceptionally tractable.

TABLE 64

DENSITY, PRESSURE, GRAVITY AND ELLIPTICITY WITHIN THE EARTH

Legendre's law: $\rho = \rho_0$

(ν = 144.° 185 = 2.51650 radians)

Distance from center r/a	Density ρ, gm./cm^3	Pressure p, bars	Gravity g, gals.	Reciprocal of ellipticity $1/e$
Center 0.0	11.2	3.2×10^6	0	373
0.1	11.0	3.1	197	372
0.2	10.7	2.9	387	370
0.3	10.1	2.6	563	366
0.4	9.4	2.18	717	362
0.5	8.6	1.83	845	355
0.6	7.4	1.26	941	347
0.7	6.2	0.86	1004	337
0.8	5.0	0.49	1032	326
0.9	3.8	0.20×10^6	1026	311
Surface 1.0	2.6	....	982	297

Table 63 (Bullen), on the other hand, satisfies seismological data also and is therefore to be preferred. In spite of the different premises the figures in the two tables resemble one another closely. Bullen does not give the value of gravity within the earth. The data in the last column of Table 63 are based on the first paper[13] by Bullen and have been calculated by Benfield.[14] It will be seen that gravity increases within the earth till the depth is about 0.6 of the radius (core). Its value is then about $2\frac{1}{2}$ per cent greater than at the surface. The rate of increase just below the outer surface is the normal rate in free air just outside the surface *minus* $4\pi k\rho$, or, in gals per meter,

$$0.0003086 - .0000838\rho_1.$$

The ellipticity of the internal strata of the earth affects the paths of earthquake waves, but our present knowledge is not sufficient to enable us to decide between one set of ellipticities and another. It is not to be expected that the paths within the earth, assuming them to be of the brachystochrone type, will be plane curves. Even on the surface of an ellipsoid the brachystochrone curve for uniform velocity is the geodesic line, and this is not a plane curve. The problem is one for the calculus of variations.

Since the foregoing paragraphs were written, the notes of de Sitter on certain interrelated astronomical data, including the figure of the earth, have been completed and published by Brouwer.[15] This memoir aims to obtain by adjustment a consistent set of values of the quantities considered. In deriving the ellipticity of the earth the quantity $(C - A)/C$ is considered as very accurately determined by means of the luni-solar precession of the equinoxes. This quantity is corrected by the so-called "geodesic precession," or relativity effect, a quantity that naturally did not enter into earlier determinations.* The "geodesic precession" is close to the border line between negligibility and non-negligibility. The assumption of fluidity is made and the value derived for the reciprocal of the ellipticity is

$$1/e = 296.753$$

as against 297 for the International Ellipsoid.

The approximation involved in (78) is closely fulfilled by the assumptions of de Sitter and Brouwer, for which reference must be made to their article. Whether the reciprocal of the earth's ellipticity, $1/e$, is absolutely determined from $(C - A)/C$ or not depends on the number of disposable constants in the assumed law of density, but as (78) shows, fluidity being always taken for granted, it makes surprisingly little difference what law of density is assumed. The quantity $(C - A)/C$ is known with considerable accuracy.

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CHAPTER XIV

THE ELASTIC CONSTANTS IN THE INTERIOR OF THE EARTH

B. Gutenberg

(Professor of Geophysics, California Institute of Technology, Pasadena)

The elastic properties of a purely elastic homogeneous body are determined by two moduli: the bulk modulus k which controls the change in volume under a pure compression, and the coefficient of rigidity μ which determines the change of form under a pure shear. In addition, the following quantities are used frequently: Young's modulus of elasticity E, Poisson's ratio σ and Lamé's constant λ. Each two of the preceding constants determine the changes in volume and form of a homogeneous body in a purely elastic process. They are connected by the following equations, which also exhibit their functional relation to the velocities V of longitudinal and v of transverse waves and the density ρ (see also Chap. IV):

$$E = \frac{\mu}{\lambda + \mu}(3\lambda + 2\mu) = 3k(1 - 2\sigma) = 2\mu(1 + \sigma) = \frac{k}{\frac{k}{3\mu} + \frac{1}{9}} = 2\rho v^2(1 + \sigma) = 3\rho\frac{V^2 - 2v^2}{\left(\frac{V}{v}\right)^2 - 1}. \tag{82}$$

$$\sigma = \frac{\lambda}{2(\lambda + \mu)} = \frac{3k - E}{6k} = \frac{1 - \frac{2\mu}{3k}}{2 + \frac{2\mu}{3k}} = \frac{1}{2}\left[1 - \frac{1}{2\left(\frac{V}{v}\right)^2 - 1}\right]. \tag{83}$$

$$\mu = \frac{1}{2}\frac{E}{1 + \sigma} = \frac{3k}{2}\frac{1 - 2\sigma}{1 - \sigma} = \frac{3}{2}(k - \lambda) = \frac{\lambda(1 - 2\sigma)}{2\sigma} = \rho v^2. \tag{84}$$

$$k = \frac{1}{3}\frac{E}{1 - 2\sigma} = \frac{2\mu}{3}\frac{1 + \sigma}{1 - 2\sigma} = \lambda + \frac{2}{3}\mu = \rho V^2 - \frac{4}{3}\mu = \rho\left(V^2 - \frac{4}{3}v^2\right). \tag{85}$$

$$\lambda = \frac{\sigma E}{(1 + \sigma)(1 - 2\sigma)} = 3k\frac{\sigma}{1 + \sigma} = k - \frac{2}{3}\mu = \frac{2\sigma\mu}{1 - 2\sigma} = \rho V^2 - 2\mu = \rho(V^2 - 2v^2). \tag{86}$$

In fluids ($\mu = 0$) we find

$$E = 0. \qquad \sigma = \tfrac{1}{2}. \qquad \lambda = k = \rho V^2. \tag{87}$$

The occurrence of elastic processes presupposes the action of stresses. A stress is the amount of a force per unit of the area upon which the force is acting. Usually not the stress itself, but stress components are used. The component perpendicular to the surface is called *normal stress*, and the component along the surface is called *tangential stress*. If the normal stress acts toward the surface, it is a *pressure;* in the opposite direction, it is a *tension*. The effect of a pressure may be a *compression;* the effect of a tension, a *dilatation;* the effect of a tangential stress, a *shear*.

The classical theory of elasticity postulates that the strains are uniquely determined by the stresses and inversely. Internal friction, plasticity, creep and similar phenomena are assumed to be absent. Effects of such phenomena on various processes are discussed in Chap. XV. In the classical theory it is assumed, besides, that the strains are so small that their squares and products may be neglected in the stress-strain equations. Attempts to extend the theory of elasticity to the case of finite strains have been made by Murnaghan[19] and by Birch.[20]

THE RIGIDITY

The coefficient of rigidity is proportional to the tangential stress that is necessary to produce a given distortion (shear). It is a constant to be used only in purely elastic processes and, therefore, is not to be applied where plastic changes are involved. Thus, contrary to a widespread belief, neither subcrustal flow nor the spreading of continents depends in any way on the rigidity. Its main importance is in its connection with elastic shear waves. In most rocks at the surface of the earth it is of the order of 3×10^{11} dynes per square centimeter. It decreases with temperature and increases with pressure (see Chap. IV).

As it is probable that a large part of the mantle of the earth consists of vitreous material, determinations of the rigidity in such material and especially of the differences between the crystalline and the vitreous form at various temperatures and pressures are of great importance. Unfortunately, data are very scanty, and, thus far, there are none on the rigidity of vitreous materials under high pressure at temperatures above the melting point.

Detailed information on the rigidity μ in the interior of the earth depends upon observations of the velocity v of transverse waves and upon results as to the density ρ, as $\mu = \rho v^2$. The velocity v is known with sufficient accuracy for the mantle of the earth; the data on the density are less certain, especially for the deeper parts of the mantle.

No transverse waves through the core have been identified with such a degree of probability that they can be used with confidence to determine the rigidity in the core. In some instances waves have been observed which possibly could be transverse waves through the core (see Chap. X), but they could just as well be waves of other types. As there is no known reason why the transverse waves through the core should be very much smaller than the longitudinal waves and since the transverse waves through the mantle usually have noticeably larger amplitudes than the longitudinal waves and most seismologists working on this special problem have failed thus far to identify transverse waves through the core, the proper conclusion seems to be that their existence thus far is very doubtful. Doubtless they do not have a focal point, as do direct longitudinal waves through the core and reflected waves through the core, or waves that have traversed the mantle once as transverse waves and the core as longitudinal waves.

Thus, the failure, up to this time, to establish transverse waves within the core makes it impossible to calculate values for the rigidity inside the core and makes it probable that it is very small, possibly even practically zero.

In the earth's crust the findings concerning the velocity of shear waves in a given layer vary somewhat (Table 36); besides, the density is known only approximately. Thus, the figures of Table 65 represent only average values; for in sediments the velocity v varies from sample to sample and the data for the deeper layer are uncertain owing to the fact that the density is not known accurately.

TABLE 65

COEFFICIENTS OF RIGIDITY

Material	Approximate Coefficient of Rigidity, Dynes/sq. cm.
Alluvium near the surface	0.1×10^{1}
Alluvium at a depth of 2 km	1×10^{1}
Tertiary sandstone at a depth of 2 km	2×10^{11}
Very old sediments at a depth of 2 km	4×10^{1}
Granitic layer in continents	3×10^{1}
Deeper continental layers	4×10^{1}
Pacific basin at a depth of 5 km	5×10^{1}
Depth of 50–100 km. everywhere	$6\frac{1}{2} \times 10^{11}$
Depth of 200 km	7×10^{11}

For greater depths, the results are given in Fig. 25 (curve 3).

Although the velocity of transverse waves combined with the results as to the density gives the rigidity as a function of depth, there are some other phenomena that permit the calculation of the rigidity of the

earth as a whole. To get results regarding the variation with depth, the various types of observations must be combined and a functional relationship between rigidity and depth must be assumed which contains as many variables as there are observed constants.

The two most important sources for information regarding the rigidity of the earth as a whole are the movements of the poles and the tides of the earth's body, as well as of the ocean; these have been discussed by W. D. Lambert (vol. II of this series, "The Figure of the Earth," pages 245–277 and 68–80, respectively). The literature on the application of the results to the calculation of the rigidity of the earth is very extensive (see references *op. cit.*).

The observations on tides and the movements of the poles furnish three quantities h, k and l. They are connected with the rigidity μ of a homogeneous incompressible body of the size of the earth (radius R) and its mean density ρ and rigidity μ in the following way:[13]

$$h = \frac{5f}{2f + 1}, \qquad k = \frac{3f}{2f + 1},$$

where

$$f =$$

In general, these three quantities are defined only for deformations expressible by second-degree spherical harmonics. This includes, however, all types of deformations that have actually been calculated from observations of tides, deflections of the vertical, changes in gravity and variation of latitude. Furthermore, the definitions apply to the free surface only.

h and k serve as a means of comparing the resulting effects with the forces producing them. h is the ratio of the actual yielding, due to the earth tide, to the theoretical equilibrium tide, measured from the undeformed surface of the earth. The yielding of the earth produces a potential kW in addition to the potential W of the long-period tidal forces. This defines k. The equilibrium height of the ocean tide is $\frac{(1 + k)W}{g}$. The tide in the earth's body is $\frac{hw}{g}$. Their difference which we observe in the free ocean from long-period tides is $\frac{(1 + k - h)W}{g}$. As we know $\frac{W}{g}$ from theory, $1 + k + h$ can be calculated. The corresponding ratio (effect of the yielding to the effect on an ideal earth) for the variation in gravity is given by $1 + h - \frac{3}{2}k$.

The horizontal surface displacement due to the tide involves a third quantity l. If the equilibrium tide E is given by cS, where S is a

general surface spherical harmonic of degree two and c is a constant, then the surface displacement in the meridian is given by $cl\ (\partial S)/(\partial \rho)$ and in the prime vertical by $\frac{(cl)}{(\cos\varphi)}\frac{(\partial S)}{(\partial\lambda)}$ (φ = latitude, λ = longitude). The direction of the plumb line is affected by the quantity l. The ratio of the value found on the yielding earth to the value for an ideal earth is given by $(1 + k - l)$.

The study of the movements of the poles gives k.

$$= \left(1 - \frac{t_E}{t_C}\right)\left(\frac{2\alpha}{m} - 1\right). \tag{89}$$

t_E is the Eulerian period of the movements of the poles for an unyielding earth (303 days), t_C is the actually observed period, corrected for the mobility of the oceanic waters, α is the flattening of the earth ($\frac{1}{297}$) and m the ratio of the centrifugal force of rotation to gravity at the Equator.

Thus, the observations of various phenomena give $(1 + k - h)$, $(1 + h - \frac{3}{2}k)$, $(1 + k - l)$ and k. Unfortunately, most of these quantities cannot be determined exactly, as the ocean waters affect the observed data and as the ocean tides, especially, produce effects on the continents that are superimposed on the body tides. Moreover, most of the quantities have not been measured accurately enough, partly because observational methods do not allow the accuracy desired, partly because the quantities show large variations due to changing disturbing effects that are of the same order as the desired quantity, so that long series of observations are needed. For most elements, the observations have not been made for a long enough time interval. The quantity l has been much neglected, although its importance has been pointed out by Lambert (*op. cit.*, page 78).

The first attempts to find the rigidity of the earth were made under the assumption that the earth is homogeneous. In this case equation (88) can be used. Lambert (Ref. 13, page 18) has calculated corresponding values of the quantities involved:

μ, dynes/sq. cm.	h	k	l	$1 + k - h$	$1 + k - l$	$1 + h - \frac{3}{2}k$
1×10^{11}	1.96	1.18	0.59	0.22	1.59	1.20
2×10^{11}	1.61	0.97	0.48	0.36	1.48	1.16
4×10^{11}	1.19	0.71	0.36	0.52	1.36	1.12
8×10^{11}	0.78	0.47	0.23	0.69	1.23	1.08
12×10^{11}	0.58	0.35	0.17	0.77	1.17	1.06
20×10^{11}	0.38	0.23	0.12	0.85	1.12	1.04

Darwin[1] found that the long-period ocean tides have amplitudes of about $\frac{2}{3}$ of the theoretical value for an unyielding earth. Although it is doubtful whether the theory can be applied to long-period ocean tides—short-period ocean tides are affected by many other factors—we shall assume that this is approximately $1 + k - h$. Then we find from the table, supposing a homogeneous earth, a rigidity of about 8×10^{11} dynes per square centimeter. Similar values have been found from other data, *e.g.*, k has been found to be about $\frac{1}{4}$ from the movement of the poles; from such conclusions came the well-known statements that the earth as a whole is about as rigid as steel (8×10^{11} dynes per square centimeter), or, with better approximation, about twice as rigid as steel if the observations of the tides are corrected for the mobility of the ocean waters and the observed value of k is considered.

However, to find the actual rigidity of the earth, it must be considered that the effect of the rigidity at various depths is quite complicated. As only a few quantities are known, relatively simple forms for the rigidity as a function of depth must be assumed. Because of this fact, combined with the difficulty of getting good numerical data for the quantities used for the calculations, this group of methods has not yet furnished us with accurate results concerning the rigidity inside the earth's core, where the more accurate method using the velocity of transverse waves cannot be applied.

The first to suppose that the rigidity throughout the earth is not constant seems to have been Herglotz,[3] who assumed that there is one constant value for the rigidity in the mantle and another constant value in the core. The next step was taken by Schweydar[6] who supposed that both the density ρ and the rigidity μ in the earth can be expressed as a function of depth by an expression of the form $a - br^2$, where a and b are constants (different for the rigidity and the density) and r the radius of the layer divided by the radius of the earth. Using the movements of the poles with Chandler's period, he found, under the assumption just mentioned, that the rigidity increases from about 3×10^{11} dynes per square centimeter at the surface to about 30×10^{11} dynes per square centimeter in the center of the earth.

Similar suppositions were made by Hoskins.[9] The density law that he assumed is similar to that of Schweydar.

Schweydar	Hoskins	
$\rho = 10.1(1 - 0.764r^2)$.	$\rho = 11.054(1 - \frac{5}{6}r^2)$.	(90)

As the bulk modulus affects the results, Hoskins made various assumptions regarding this, too. The best approximation which he has con-

sidered supposes that Lamé's constants λ and μ, are equal, which corresponds fairly well to the actual conditions in the mantle of the earth. However, the density assumed for the surface layers ($\rho = 1.8$ for $r = 1$ from the equation for ρ given above) is too small. By using the observed data on the tides, he found for the rigidity under his assumptions

$$\mu = 16.95(1 - 0.85r^2), \tag{91}$$

which corresponds to an increase from his assumed value of about $2\frac{1}{2} \times 10^{11}$ at the surface to about 17×10^{11} dynes per square centimeter at the center, increasing, according to his reasoning, first faster, then more slowly, and finally approaching the value at the center asymptotically.

A. Prey[10] approached the problem in a different way. He extended the theory given by Schweydar and tried to combine data found from the tides of the earth with those found from the movements of the poles, hoping that in this way he could suppose a function containing one more variable. However, he found that the two phenomena yield equations that do not differ by more than the possible errors in the observations. This result is very important since it shows that the values for the rigidity found in completely different ways agree; however, it makes it impossible to get more information on the law that corresponds to the change of the rigidity with depth. In his first paper, Prey used data based on the combined results from observations of tides and polar movements, and in the second he used the period of the movements of the poles only but extended the theory by considering some important terms of the second order.

For the density, Prey assumed the same law as Schweydar (see above). For the rigidity μ he made two different assumptions, the first of which agrees with those of Schweydar and Hoskins: increase of μ from the surface to a maximum in the center of the earth. Assuming a value of 2.7×10^{11} for the rigidity at the surface, he found the following as the most probable law for the rigidity inside the earth, under the suppositions just mentioned:

$$\mu = 16 \times 10^{11}(1 - 0.83r^2). \tag{92}$$

This equation agrees well with the results of Hoskins, found in a wholly different way. The value adopted for the rigidity at the surface almost corresponds to that of the granitic layer in the continents. As the value is about twice as high at the bottom of the Pacific Ocean and under the continents at a depth of about 40 km., the present writer suggested a calculation to Prey (communication by letter)

under the assumption that the rigidity is 6×10^{11} dynes per square centimeter near the surface. The corresponding law for the rigidity under otherwise equal assumptions is about

$$= 15 \times 10^{11}(1 - 0.6r^2). \tag{93}$$

The values corresponding to these results have been plotted in Fig. 25, curve 1. This curve, as well as the others, found under the supposition that the rigidity increases with depth to a maximum value at the center of the earth, disagrees completely with the values found from earthquake waves in the mantle, where they are by far too low. Since, in addition, no transverse waves through the core have

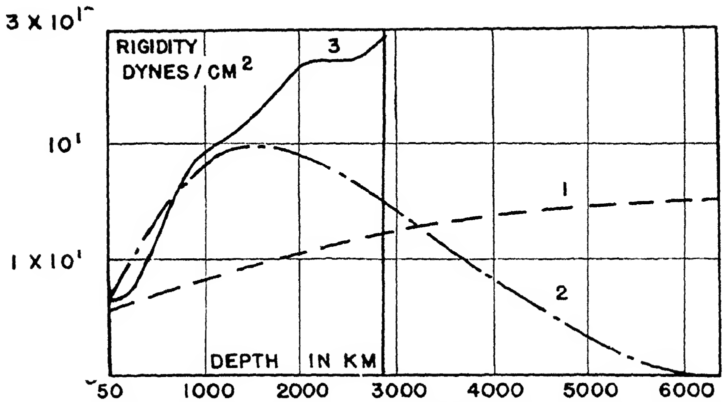

FIG. 25.—Rigidity in the interior of the earth as a function of depth. Curves 1 and 2 are based on observations of the body tides of the earth and on polar movements, curve 3 on the velocity of transverse earthquake waves and the density in the earth.[18]

been found thus far with properties corresponding to those of longitudinal waves through the core, it seems probable that the rigidity inside the core is very small. For this reason, Prey has assumed a second type of law for the change of rigidity with depth:

$$\mu = Ar^2(1 - Br^2).$$

For the surface, we must again adopt the value of μ. With increasing depth, the rigidity first increases to a maximum and then decreases to zero toward the center of the earth.

Prey has carried out the calculations by combining the results from the tides and the polar movements and by using the results of the polar movements alone, employing in this latter case the equations developed in his second paper. He again assumed the value of 3×10^{11} dynes per square centimeter for the rigidity near the sur-

face but by letter communicated in addition the results, supposing that the rigidity near the surface is 6×10^{11}. The results are as follows:
From tides and polar movements,

$$\mu_0 = 3 \times 10^{11}: \qquad \mu = 60 \times 10^{11} r^2(1 - 0.95r^2). \tag{94}$$
$$\mu_0 = 6 \times 10^{11}: \qquad \mu = 70 \times 10^{11} r^2(1 - 0.91r^2). \tag{95}$$

From polar movements alone,

$$\mu_0 = 3 \times 10^{11}: \qquad \mu = 100 \times 10^{11} r^2(1 - 0.97r^2). \tag{96}$$

The differences among the various curves are relatively small. The maximum value of the rigidity occurs in all cases at depths between 1,500 and 2,000 km. and is of the order of 20×10^{11} dynes per square centimeter.

The curve corresponding to the second equation has been plotted in Fig. 25, curve 2. Considering the fact that the form of the curve is given by the assumption, it is a good approximation to the curves found from the velocity of transverse waves; it indicates that the rigidity in the core is very small.

An attempt to calculate the rigidity of the earth's crust from the tilt produced by the passing of a low-pressure area has been made by H. Lettau.[14] As has been mentioned in Chap. VIII, the maximum tilt i in seconds of arc is given by $10^8 h/\mu$, from which we find $\mu = 1.36 \times 10^8 H/i$ dynes per square centimeter, where H is the change in air pressure in millimeters mercury. From observations taken near Leipzig, Lettau found $i = 0.004''$ for $H = 1$ mm. Hg., which gives $\mu = 0.3 \times 10^{11}$ dynes per square centimeter. A similar value for μ would explain the semidiurnal wave recorded by pendulums, under the assumption that this is a combination of a solar tidal wave and the corresponding semidiurnal air-pressure wave. The value of 0.3×10^{11} would correspond almost to the rigidity of alluvial layers, as this has been found from elastic waves as well as in laboratories. However, it is only about one-tenth of the value found for granite. For this reason Lettau considers the possibility, mentioned before by Schweydar (Ref. 6, page 41) and by Prey,[10] that very slow movements, as in body tides, indicate a smaller coefficient of rigidity than much faster movements in earthquake waves. Further investigations are needed to decide this question.

A method for finding the rigidity inside the core would be to assume for the mantle a curve found from the velocity of the transverse waves and from an adopted curve of the density, *e.g.*, curve 3 in Fig. 25, and then to calculate the rigidity in the core (which can be assumed

to be constant as a first approximation) from the tides or polar movements. However, this problem offers great mathematical difficulties which thus far have prevented its solution. An approximation to it is based on the assumption that the rigidity has one constant value in the mantle and another in the core. The solution has been given by Herglotz[3] and applied by Jeffreys,[11] who assumed an "average" rigidity for the mantle from the data found from earthquake waves and then tried various assumptions on the rigidity of the core as to their effect on the tides and polar movements.

Jeffreys found that the results

> . . . definitely imply that the core is less rigid than the shell absolutely, and not merely in comparison with its density. The results for a liquid core are a little too high, but it appears probable that they would be appreciably reduced if we allowed for variations of density and rigidity within the two layers. A liquid core would then make it possible to reconcile the tidal and seismological data; a core with a rigidity related to its bulk-modulus (about 10^{13} dynes/cm.^{2}) in any ratio admissible for a solid would be entirely impossible.

Under the assumption that the rigidity of the core is zero, L. Rosenhead[12] made a detailed theoretical investigation on the tides and found a good agreement between the observations and the theory.

From all the evidence available it seems probable that the rigidity of the core is relatively small. It does not follow from the observations that it is so small that we should call the material fluid; but it is possible that it is. In any case, the assumption that it is zero is at least a good first approximation for many problems involving the rigidity of the earth. It is not impossible that transverse waves through the core with a relatively small velocity may be recognized in the future, but it is very unlikely that the ratio of the transverse and longitudinal velocities is of the same order in the core as in the mantle. The hypothesis that, for stresses which act over long periods in the same way (polar movements, tides), the rigidity may be small while it is large for the short-period transverse waves is supported neither by theory nor by observations.

The finding that the rigidity inside the core is close to zero would suggest that the core is not solid. Therefore, the conclusion has frequently been drawn that the boundary of the core is a boundary between molten material inside the core, and crystalline material or vitreous material below its melting point, outside the core. However, there is no reason why material could not have almost equal rigidity above and below its melting point at the high pressure of more than 1

million atmospheres. On the other hand, the core probably consists of metal, mostly iron (see Chaps. V and IX), whereas the material of the mantle, even near the boundary of the core, most likely is a type of rock or ore. It is not impossible that both materials are at a temperature above their melting point near the boundary but that the rigidity of the metal under the conditions there is a few orders of magnitude smaller than that of the stony material in the deepest part of the mantle. A factor of the order of magnitude of 10^{-2} could possibly be sufficient to explain the observations of the tides and the movements of the poles. This would correspond to the ratio between the rigidity of granite and young sediments at the surface, where ratios as large as $\frac{1}{500}$ are known for solids. On the other hand, we do not know the properties of materials under such high pressures and temperatures, and changes unknown to us may produce relatively small rigidity combined with very little compressibility. Teller[15] has pointed out that "around a million atmospheres the work of compression becomes great enough to cause an essential change in the electronic structure. Matter in the core of the earth is in this transition-region in which all physical and chemical properties will begin to change strongly." However, conditions that have been suggested for the interior of the stars are out of the question for the core of the earth.

The question now arises whether zero rigidity in the core does not produce gravitational instability. This problem has been investigated by E. Meissner[16] and H. Jeffreys,[11] using the theoretical investigations of Lord Rayleigh and A. E. H. Love.[17] Meissner supposed that the two Poisson constants μ and λ are equal and found that if this is true the rigidity in the center must be at least 15×10^{11} dynes per square centimeter to prevent radial instability. Jeffreys pointed out that the bulk modulus and, therefore, λ have been found to be large in the core. He came to the conclusion that incompressibility would maintain stability for radial displacements, even if the rigidity were zero everywhere, and that there is no reasoning which proves that the earth's core cannot be truly fluid.

THE COMPRESSIBILITY AND THE BULK MODULUS

The bulk modulus k is proportional to the change in pressure Δp which is needed to change a given volume v by a certain amount Δv. $k = -(\Delta p/\Delta v)v$. The larger the bulk modulus, the smaller the compressibility.

Though determinations of the bulk modulus in laboratories are more easily made and data are more plentiful (see Chap. IV) than those for the rigidity, there are fewer possibilities for its determination in the

interior of the earth. The only method that has been used up to the present depends on the velocities of longitudinal waves (V) and transverse waves (v) in combination with the density ρ.

(97)

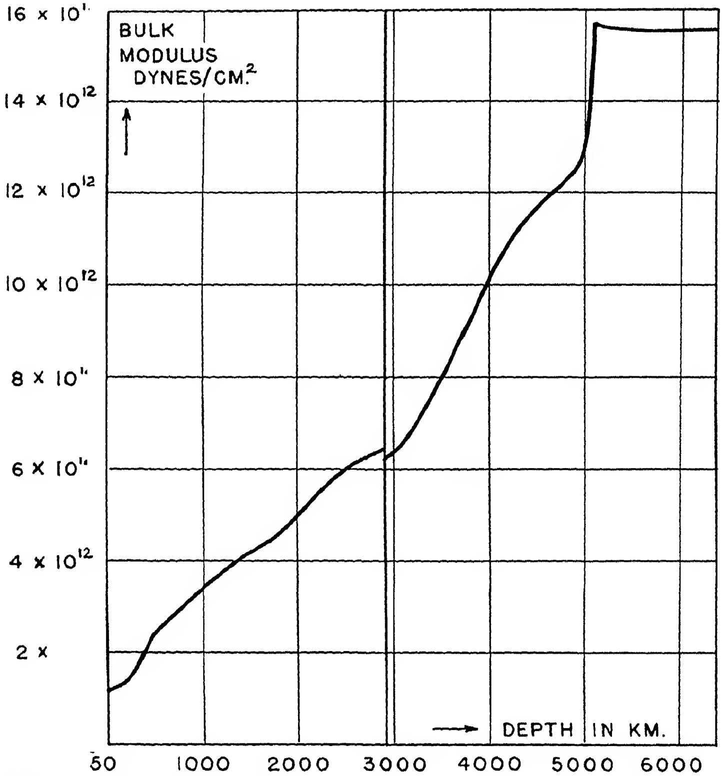

FIG. 26.—Bulk modulus in the interior of the earth as a function of depth, based on the velocity of longitudinal and transverse earthquake waves and the density of the earth.[18] It has been assumed that inside the core the velocity of the transverse waves is very small and negligible.

As far as data are available, the values determined in this way seem to agree within small limits with those found by static methods in laboratories.

In the granitic layer of the continents the bulk modulus found from the velocities of seismic waves and the density is about 5×10^{11} dynes per square centimeter; in the Pacific Ocean at a depth of a few

kilometers it is probably not far from 10×10^{11} dynes per square centimeter. At depths of between 50 and 100 km. it is about 12×10^{11} dynes per square centimeter everywhere. For the greater depths, the probable values are plotted in Fig. 26.

For the core, zero rigidity has been assumed. If the rigidity there is not zero but small, the calculated bulk modulus would be slightly less than that given in Fig. 26, but the difference would be within the limits of error.

From Fig. 26 it follows quite clearly that, unlike the rigidity, the bulk modulus does not change much at the transition from the mantle to the core. The material inside the earth approaches incompressibility more and more toward the center of the earth.

YOUNG'S MODULUS OF ELASTICITY

Most experiments on the elastic constants concern Young's modulus. In geophysics and especially in seismology, Young's modulus does not play an important role, as it is usually calculated from the rigidity and the bulk modulus. Young's modulus E is the force that must be applied at the end of a cylinder to double its length, under the supposition that the change in length is proportional to the force, regardless of its magnitude (Hooke's law). This supposition is correct within the limits of error in the treatment of most geophysical problems, since in such cases the forces and elastic changes are usually small.

Formulas for the calculation of E are to be found at the beginning of this chapter. It is of the order of 2×10^{11} dynes per square centimeter in sandstone, about 7×10^{11} in the granitic layer of the continents, about 12×10^{11} beneath the bottom of the Pacific Ocean at a depth of a few kilometers, about 15×10^{11} everywhere at a depth between 50 to 100 km. It increases with depth in a way similar to that of the bulk modulus and the rigidity and has its maximum of about 10^{13} dynes per square centimeter in the mantle near the boundary of the core. Inside the core it is very probably considerably less than 10^{12} dynes per square centimeter, possibly near zero.

POISSON'S RATIO

If a cylinder is strained by forces acting at both ends in opposite directions along the axis, the diameter changes as well as the length. The ratio of these two changes is called *Poisson's ratio.* Its maximum is $\frac{1}{2}$ for a material without rigidity, as well as for an incompressible material, as it depends only on the ratio of the rigidity to the bulk

modulus. Its minimum is zero for an absolutely rigid material (or for one with zero bulk modulus). For most rocks it is near $\frac{1}{4}$.

For the interior of the earth Poisson's ratio σ can be found from the ratio a of the velocity of longitudinal waves to that of transverse waves.

$$\begin{matrix} \frac{1}{2}a^2 - 1 \\ a^2 - \end{matrix} \tag{98}$$

The values of σ found for the granitic layer are slightly less than $\frac{1}{4}$; in the deeper continental layers they are about $\frac{1}{4}$, and the values for the deeper parts of the mantle are slightly greater, with a maximum of about 0.3 near the core. Inside the core, Poisson's ratio seems to be close to $\frac{1}{2}$.

If Poisson's ratio is 0.27, which is a good approximation for most parts of the mantle, the following approximate equations result for the other elastic constants and the velocities of elastic waves (symbols as before):

$$E = 0.8\rho V^2 = 2.54\rho v^2 = 2.54\mu = 1.4k \qquad V:v = 1.78 \qquad \mu = 0.54k. \tag{99}$$

SUMMARY

The rigidity increases inside the earth with increasing depth from about 10^{10} dynes per square centimeter in recent sediments to about 3×10^{11} dynes per square centimeter in the granitic layer of the continents, about $6\frac{1}{2} \times 10^{11}$ at a depth of 100 km., about 10^{12} at a depth of a few hundred kilometers, 2×10^{12} between 1,000 and 1,500 km. and about 4×10^{12} dynes per square centimeter in the mantle near the core at a depth of 2,900 km. Inside the core it is noticeably less, probably less than 10^{11} dynes per square centimeter, possibly of a much smaller order of magnitude. Even if the core has so small a rigidity and viscosity that it behaves as a fluid, it would be stable on account of the high bulk modulus.

In the crust and the mantle the bulk modulus is about twice the coefficient of rigidity but continues to increase inside the core slightly beyond 10^{13} dynes per square centimeter. Young's modulus of elasticity is about $2\frac{1}{2}$ times the coefficient of rigidity and decreases suddenly at the boundary of the core in a way similar to the rigidity. Poisson's ratio seems to increase, not quite regularly, with depth inside the mantle from about $\frac{1}{4}$ in the crust to about 0.3 near the core; inside the mantle, in general, 0.27 is a good approximation. Inside the core it is probably not far from $\frac{1}{2}$.

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CHAPTER XV

VISCOSITY, STRENGTH AND INTERNAL FRICTION IN THE INTERIOR OF THE EARTH

B. GUTENBERG

(*Professor of Geophysics, California Institute of Technology, Pasadena.*)

If a very small stress is applied to a rock in the earth's crust, the strain corresponds very closely to the equations for purely elastic materials, especially Hooke's law. If the stress increases, the agreement becomes less good, the material changing less than is required by Hooke's law. If shearing stresses continue to increase, a point is finally reached where the material continues to change its shape even if the stress difference is not increased. This critical stress difference is defined as the *strength* of the material. If the stress difference is larger than the strength, the material undergoes *plastic flow.* The ratio of the stress difference to twice the rate of shear during plastic flow is called *coefficient of viscosity* or simply *viscosity.* It is usually assumed to be a constant, but the experiments show that this is only a very rough approximation.

If the stress difference increases still more, a rupture will occur. The corresponding stress difference is the *breaking strength.* Sometimes the material breaks before it begins to flow. Care should be taken to avoid confusing the two meanings of the term *strength.* Although the expression *strength* is generally understood to mean strength against viscous flow, some writers use it as synonymous with breaking strength. However, these two quantities seem to have no clear relationship to each other. The same seems to be true for the relationship between rigidity and viscosity. Wax is very plastic but nevertheless rigid; rubber has relatively small rigidity but very little plasticity.

As has been mentioned, Hooke's law provides only an approximation of the elastic processes. The observations can be represented by an equation of the form

where S is the distortion (strain), F is the tangential stress, μ is the

rigidity, ϑ and β are constants of the material and f is a function of several variables, as amplitude, density, etc., involved in the process.

The equation $S = F/\mu$ represents Hooke's law. If we add the second term, we consider a retarding effect which frequently is identified with *internal friction*. However, to get accurate results, at least all the terms written down in Eq. (100) must be considered. As they make a theoretical investigation of most problems impossible, most of them are usually omitted.

This fact, as well as the confusion due to the use of such expressions as *creep, elastic afterworking, internal friction* or *viscosity* for different processes, complicates the problem. Certain experiments show deviations from Hooke's law, but no afterworking; in others the final strain follows Hooke's law with good approximation only after enough time has elapsed to allow elastic afterworking to produce the final strain. Other processes show gradual macroscopic changes which frequently are called creep. Some of these and other deviations from Hooke's law possibly have the same cause; internal friction probably is one of them. On the other hand, Eq. (100) of the process that we have called internal friction covers various other minor processes.

What we get in this way is, at best, a compromise. Even the first approximations now used make it practically impossible to solve the equations in many problems. Another difficulty arises from the fact that very few experiments have been made, and those which have been made involve very few materials.

NONLINEAR ELASTICITY AND CREEP

The most complicated problems in this field are those in which the stress differences are so large that the pure elastic theory no longer holds but in which the differences are still below the viscous strength as well as below the breaking strength. Further complications are added if the body is inhomogeneous. Attempts to describe the processes mathematically have been made by Schlechtweg.[1] His complicated *nonlinear elasticity law* is a better approximation than the equations based on Hooke's law. Only for extremely high pressures are the differences between observations and the new law noticeable. Other investigations have been made by Phillips.[37]

Jeffreys[33,36] considers elastic afterworking as the main cause of imperfect elasticity. Internal friction possibly accounts for it. He believes that usually, owing to the incompleteness of the experiments, there is no way to decide in which cases creep is due to elastic afterworking and in which to plastic flow. He considers elastic afterworking as due to imperfection of crystal structures and points out[2]

that it is absent from perfect crystals. He gives the following equation for elastic afterworking:

$$\mu\left(\frac{d}{dt} + \frac{1}{\tau'}\right)S = \left(\frac{d}{dt} + \frac{1}{\tau}\right)F, \tag{101}$$

where $\tau' > \tau$ are constants. τ' gives the time scale of the process, τ'/τ its amount. The equation may be written[36]

$$\mu S = F\left[1 + \left(\frac{\tau'}{\tau} - 1\right)(1 - e^{-t/\tau'})\right]. \tag{102}$$

Thus far imperfect elasticity has not played an important role in geophysics. Moreover, the observation of earthquake waves has not yet led to results that could be explained only by assuming imperfect elasticity, although some phenomena may be due to this.

There is another group of processes which are of the type of plastic flow but which occur at stress differences much less than strength. The presence of this group, as well as the fact that they do not follow the theory of plastic flow, leaves no doubt about systematic differences between the two types of flow. The type considered here is usually called creep. Apparently, no clear definition of the term has been given, although in publications it is usually described as "well known." It is a very slow movement under stress differences that are below the strength. Up to the present, no limitation has been made concerning the type of motion, whether gradual, in small steps or a combination of both.

Creep may be defined as relatively slow movements along surfaces with dimensions larger than single crystals. The closing in of rocks in mines, the movements of sediments along slopes and gradual movements along faults may be due to creep. Mud flows, however, are closer to viscous flow. In sliding areas large amounts of rock frequently move downward with a speed that does not vary much from day to day. A sensitive seismograph on a slide in Los Angeles, in which about 2 million tons of rock were sliding down about 2 cm. per day for many days, did not record any sudden movements during a test of several hours.

Geodetic measurements in Japan have shown that along most of the active faults the movements between earthquakes continue in the same direction. An example of this type in California has been reported by Koch.[3] In the Buena Vista Hills oil field, Kern County, relatively large movements, about 4 cm. per year, have been observed for a number of years without any felt shock or any sudden movement

along a fault. "The fault probably is the result of differential slipping between beds on the north flank of the structure induced by current folding." Figure 27 shows the bending of pipe lines (front, center and about one-third from the left in the background below the arrow). The fact that casings of wells are sheared off at various depths depending on their location has made it possible to calculate that the fault dips approximately 25° to the left (higher surface) in the figure. The block to the left is overriding the one to the right. Other gradual movements of this type seem to occur in the same region. Koch mentions that "the concrete highway, a mile south of the town of McKittrick, is being buckled up to such an extent as to necessitate

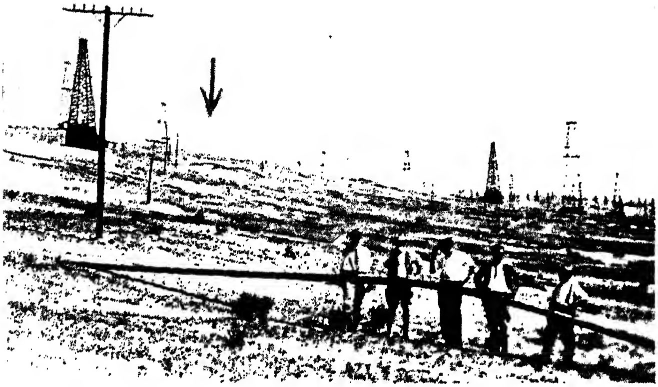

FIG. 27.—Bent pipe lines indicating faulting movements in the Buena Vista Oil Field, Calif.

repairs about once every two years." Measurements indicate a movement of about 2 cm. per year. Again, no sudden movements have been observed. In both instances, the movements have to be classified under creep. If there are sudden movements, they must be very small and occur so frequently as to produce the impression of a gradual movement.

Though for processes occurring near the surface of the earth there is usually no difficulty in discriminating between creep and plastic flow, this problem is more serious for the earth's interior.

INTERNAL FRICTION

The deviation of the observed distortion from Hooke's law may be given by the second and following terms of the right-hand side of Eq. (100). The source of this deviation is frequently called *internal*

friction, though many causes are involved. H. Jeffreys has called it *firmo-viscosity*, but he no longer considers this term correct.[33] As this book is not the proper place to introduce new words for elastic and plastic constants, the expression *internal friction* is used; however, we must keep in mind that internal friction is only a part of the processes that we describe under this name. Measurements of the internal friction in reeds indicate[52] that in this case the effect of the flow of heat back and forth across the reed is of a larger order of magnitude than that of all other causes.

To a first approximation, it can be assumed that pure changes in volume occur without internal friction, as the elements of a compressed material do not change their relative positions. According to Adams, experiments agree with this theoretical conclusion, and apparently it applies to all forms of imperfect elasticity.[33] However, internal friction delays the elastic strain due to shearing stress.

Contrary to a widespread belief, internal friction and plastic flow of solids are not related to each other. The first produces a delay in elastic shearing; the second produces a distortion in addition to that in the purely elastic process and could be called *internal sliding*, if the word *sliding* did not have a definite meaning. The word *plasticity* is used in this chapter; its main disadvantage is its common use which may include properties not desired by us.

There is no possibility of using Eq. (100) for practical purposes. Although frequently the term with the function f is larger than the term with the constant ϑ, it must be neglected, as otherwise the equations including internal friction cannot be solved. For this reason, we follow a suggestion of Sir J. Larmor to H. Jeffreys (Ref. 4, page 263) and describe the effect of the internal friction to a first (but frequently not good) approximation by the equations

$$\eta \frac{\partial S}{\partial t}, \quad (103)$$

where S = distortion, μ = rigidity, F = tangential stress and ϑ = a constant of the material. $\eta = \mu\vartheta$ must be constant also; it is called the *coefficient of internal friction*.

If the tangential stress F is constant, the displacement approaches asymptotically the value F/μ which corresponds to the displacement that would be immediately observed in the case of pure elasticity. If the stress is removed, we have $S + \vartheta(\partial S)/(\partial t) = 0$, $S = e^{-t/\vartheta}S_0$; the strain diminishes exponentially to zero. ϑ is thus the time in which the strain falls to $1/e$ of its value, if the stress is removed.

In a perfectly elastic body the strain increases immediately after the occurrence of a stress F to its full value F/μ. Conversely, if we assume Eq. (103), the initial strain is zero. However, it increases very rapidly with time:

$$S = \frac{F}{\mu}(1 - e^{-t/\vartheta}). \tag{104}$$

To apply this equation, we must choose a value of ϑ which has been derived from a process similar to the one for which the equation is to be used, as the observed values of ϑ contain, besides the theoretical values of ϑ, also the variable quantity f of Eq. (100) and thus depend on the period, amplitude and other quantities. From torsional vibrations of bars, ϑ has been found to be of the order of 10^{-4} to 2×10^{-3} sec. for many metals. In torsional vibrations of these metals, therefore, the strain must have reached about 95 per cent of its final value after the time of the order of magnitude just given. In tuning forks, this time must be still shorter. There is no doubt that this result agrees better with the physical conception of such a process than the assumption that the strain jumps to F/μ "in no time." However, we must always consider the fact that the observed values of ϑ are a function of many variables. Birch and Bancroft[6] found that the logarithmic decrement of free vibrations of a long column of granite was about 0.02, roughly independent of frequency, between 140 and 4,500 cycles per second. Thus, ϑ and η are about proportional to the period in this case. Further investigations to find the value of ϑ under various circumstances are badly needed.

As Jeffreys has pointed out, for motions periodic in time long compared with ϑ, ($\partial S/\partial t$ very small as compared with ϑ) we find $F = \mu S$; thus, for very long period motions, the effect of internal friction is small.

In fluids, the rigidity μ is near zero, but, except for ideal fluids, the internal friction remains finite. Thus, Eq. (103) reads for fluids with internal friction

$$F = \eta \frac{\partial S}{\partial t}. \tag{105}$$

As the viscosity in fluids is due to the internal friction, the constant η is frequently called *coefficient of viscosity* for fluids.

For solids, η as defined above can be found approximately for a given process, *e.g.*, from observations of the torsional vibrations of a bar (length l, radius r, period of vibrations T, logarithmic decrement of amplitudes D, moment of inertia J).[5]

$$\eta = \frac{4Jl\,D}{\pi r^4 T} \tag{106}$$

For metals, η has been found to be of the order of 10^9 poises (1 poise = 1 g. per centimeter per second. The unit is called *poise* in honor of Poiseuille who investigated the viscosity of fluids). Other authors have defined different coefficients of internal friction.[6]

As Jeffreys (Ref. 4, page 265) has pointed out, it is possible to find to a first approximation the effect of internal friction in a process by replacing in the terms involving shear the quantity $\mu f(t)$ by the operator $\mu f(t) + \eta \partial[f(t)]/\partial t$ and the quantity $\lambda f(t)$ by the operator $\lambda f(t) - \frac{2}{3}\eta \partial[f(t)]/\partial t$. The first to investigate the effect of internal friction on the propagation of elastic waves seems to have been Sezawa.[7] He supposed that both shear and change in volume are delayed by internal friction and gave the theoretical results for a number of wave forms. More detailed results have been found by Gutenberg.[5,8] For a single wave with the velocity V, the original period T_0 increases to a period T at a distance D, following to a first approximation the equation

$$T^2 = T_0^2 + \frac{a\,D}{V^3} \tag{107}$$

where a is a constant depending on the wave form and the material, especially the internal friction. In general, the waves get longer and flatter with increasing time (or distance).

Approximately, for many wave forms $\eta = \rho a/5$, where ρ = density. Thus, for the crust the coefficient of internal friction is about $\frac{1}{2}a$, for the deeper parts of the mantle about a and for the core about $2a$.

The increase with distance in the periods of waves has been investigated by Gutenberg for various types of waves. From these investigations, the following order of magnitude of the internal friction may be derived:

Type of wave	Region of path	Range of periods, sec.	a, sq. cm./sec.	η, poises
Surface waves[9]	Surface layers, depth increasing with period	1 –30	About 10^{10}	About 5×10^9
Longitudinal, transverse.[10]	Mantle, outer part	0.1 –10	Slightly less than 10^{10}	About 5×10^9
Microseisms[11]	Surface layers	2 –10	3×10^9	About $1\frac{1}{2} \times 10^9$
Artificial explosions, longitudinal waves.[12]	Sediments	0.01–0.04	About 2×10^7	About 10^7

This indicates a value of 10^9 poises for the internal friction in the earth's crust, except for the sediments, where the observations of longitudinal waves indicate a smaller value of the order of magnitude of 10^7 poises. These waves produced by artificial explosion had periods of about 0.02 sec. The magnification of the instruments decreased rapidly with periods increasing beyond 0.03 sec., so that possibly longer waves were suppressed by the instruments. Data published by Meisser[13] on the increase of periods of waves with a velocity of 259 m. per second from 0.02 sec. near the source to 0.17 sec. at a distance of 400 m., but recorded by an instrument with a free period of 0.2 sec., lead to a value of the same order of magnitude, 10^7 poises, for η. However, these waves may not have been elastic waves, but the so-called *ground roll,* which possibly is of another type of wave. Possibly, the differences are due to the differences in period; as has been mentioned repeatedly, ϑ and η depend on the period. The values of η found in the table are approximately proportional to the period; the fact that the same result has been found from laboratory tests of a vibrating column of granite has been mentioned above.

In the foregoing calculation the effect of plastic flow on the waves has been neglected, but, as will be shown in a later section of this chapter, this is very small. The general result, therefore, is that the coefficient of internal friction in the earth's crust is probably of the order of magnitude of 10^9 poises (grams per centimeter per second) for waves with periods of the order of 1 sec. and possibly less for shorter waves.

An attempt to calculate the internal friction from the loss of energy in earthquake waves has been made by H. Jeffreys (Ref. 4, page 265). Under the assumption that the effect of plastic flow is negligible, he found $\vartheta = 0.004$ sec. Supposing a rigidity μ of the order of 10^{12} dynes per square centimeter, we find in this way that the internal friction is of the order of 4×10^9 poises. As all results agree well with each other, it seems that the theory is at least a good first approximation. However, it is doubtful whether the results permit a decision regarding the change of the internal friction with depth inside the earth. The data that are available concern only the outer part of the mantle and the values for metals in the laboratory. They do not differ much; if there is a change with depth, it seems to be a small increase in internal friction, from a few times 10^9 in the rocks near the surface to about 10^{10} poises at a depth of the order of 1,000 km. The values found for ϑ for waves with small amplitudes and periods of the order of 1 to 10 sec. are of the order of magnitude of 0.003 sec., the same within the limits of error everywhere in the outer half of the mantle of the earth.

The effect of internal friction on the Chandler movement of the poles and on the body tides of the earth has been investigated by Jeffreys (Ref. 4, pages 266–267). The decrease in amplitude of the polar movement is given by $e^{-4\pi^2\vartheta t/3T^2}$, where T = period, about 4×10^7 sec., and t = time. For one revolution of the poles ($t = T$) the exponent of e is of the order of -10^{-10}, which means that the loss in energy due to internal friction with a value of ϑ as found above is insignificant. The lag in phase of the body tides of the earth due to internal friction is given by $2\pi\vartheta/T$, which is of the order of 5×10^{-7}, and the effect of tidal friction of the body tides on the rotation of the earth is only an insignificant fraction of the effect of the friction of ocean tides in shallow seas.[53]

STRENGTH AND BREAKING STRENGTH

The word *strength*, as has already been mentioned, is used for quite different properties. We define strength as the minimum distortional stress needed to produce plastic flow. It depends on the material, the temperature and the confining pressure. Breaking strength is the minimum stress needed to produce rupture under the same conditions. Under relatively small confining pressure, rupture frequently occurs at smaller stresses than plastic flow. At high confining pressures, the two events occur usually in the reverse order; the strength is smaller than the breaking strength.[14,15] There are exceptions.

The theory of both quantities is very complicated. Moreover, the breaking strength combines the effect of various types of stresses, of which one may produce the rupture in one instance, another in another instance. However, for our purposes we consider the minimum value of all.

G. I. Taylor has pointed out[16] that pure single crystals of metals and at least some compounds behave as if they possess no strength; they acquire strength by irregularities in the structure. He considers the strength as determined by the distance apart of faults in the crystals.

There are few data available on the strength of rocks; however, their agreement is good. Table 66 gives some results.

The column headed Depth indicates the approximate depth at which the confining pressure occurs. However, the strength as given in the last column does not correspond to this depth, as the temperature is much higher there, and the strength changes considerably with a change in temperature. For example, according to Welter[49], the strength of carbon steel increases from about 6×10^9 dynes per square centimeter at 0°C. to about 8×10^9 at 200°C. and then decreases to

TABLE 66

STRENGTH OF MATERIALS AT ABOUT 15°C.

Material	Author	Confining pressure, atmospheres	Depth (approximate) km.	Strength, dyne/sq. cm.
Marble	v. Kármán[14]	700	$2\frac{1}{2}$	3×10^9
Marble	v. Kármán[14]	1,700	6	4×10^9
Marble	Adams-Bancroft[17]	1,800	7	$4\frac{1}{2} \times 10^9$
Marble	Adams-Bancroft[17]	2,500	9	5×10^9
Marble	Griggs[15]	4,000	16	1×10^9
Marble	Griggs[15]	8,000	29	$1\frac{1}{2} \times 10^9$
Marble	Griggs[15]	10,000	37	About $3\frac{1}{2} \times 10^9$
Sandstone	v. Kármán[14]	600	2	$2\frac{1}{2} \times 10^9$
Sandstone	v. Kármán[14]	2,500	9	4×10^9
Granite	Adams-Bancroft[17]	1,800	7	5×10^9
Granite	Adams-Bancroft[17]	2,500	9	11×10^9
Solenhofen limestone	Griggs[15]	6,000	22	4×10^9
Solenhofen limestone	Griggs[16]	10,000	37	$4\frac{1}{2} \times 10^9$
Quartz	Sosman			1×10^9

about 3×10^9 dynes per square centimeter at 500°C. Geller found that the strength of bischofite decreases from 16×10^9 at -40°C. to 3×10^9 at 0°C. and to $\frac{1}{2} \times 10^9$ dynes per square centimeter at 100°C. According to him, the strength of rock salt under small confining pressure decreases from about 10×10^9 dynes per square centimeter at -50°C. to about 6×10^9 at 100°C. and to 10^9 at 550°C. Moreover, it appears from experiments that the strength is influenced by repeated changes in the confining pressure.

From the foregoing data we may safely conclude that near the surface of the earth the strength is of the order of magnitude of 10^9 dynes per square centimeter, or slightly more for most types of rock. With increasing depth, its value probably does not change much since the effects of the increase in pressure and of the increase in temperature balance each other until the temperature approaches the melting point. From there on, probably starting at a depth of about 40 km., we may expect a decrease in strength with increasing depth, which becomes the more rapid, the more the region of vitreous material (see Chaps. III and VII) is approached.

If we try to find the strength in the earth as a function of depth, we must consider that we can find only the minimum stress at which no plastic flow occurs and that we cannot decide what the mechanisms of this flow are. The actual differences in elevation on the earth are possible only if the strength is of the order of magnitude of 10^9 dynes per square centimeter. As this is about the value found from experiments, it is very likely that we have to deal with the same quantity in both cases and that the height of the highest mountains is limited by the strength. On the other hand, isostasy requires plastic flow at a depth of less than 100 km. under stresses of the order of 10^7 or even 10^6 dynes per square centimeter. If we accept isostasy, we must assume that the order of magnitude of the strength at a depth of 100 km., or slightly less, cannot exceed 10^7 dynes per square centimeter and is possibly less (see also Chap. III). However, this does not require that the strength at the surface and the strength at a depth of 100 km. be of the same type. What interests us first is the minimum stress at which plastic flow begins and the speed of this flow. As long as the order of magnitude of these quantities is doubtful, there is little hope of discriminating between various types of plastic flow and the corresponding strengths.

Thus far, no observations are known which contradict the assumption that the strength as we have just defined it is of the order of 10^9 dynes per square centimeter in the upper 40 km. of the earth's crust and decreases rapidly below this point to the order of 10^7 or even less. Such a type of distribution of strength with depth—relatively large values in the upper 30 km. of the crust, decreasing rapidly below that point—had already been suggested qualitatively for geodetic reasons in 1915 by Barrell.[18] In a paper published in 1925, several years after his death, he came to the more detailed conclusion[19] that the strength is of the order of magnitude of 10^9 dynes per square centimeter at the surface, increases to the order of 10^{10} at a depth of about 20 km., then decreases rapidly with depth to the order of 2×10^8 at a depth of 50 km. and continues to decrease, but more slowly to the order of 10^8 at 100 km. and of 10^7 at a few hundred kilometers. Jeffreys[33,34,35,36] believes that the strength at a depth of 600 km. is at least of the order of 10^8 dynes per square centimeter and that we need a new explanation of isostasy: he considers isostatic readjustment as due to sliding on planes of fracture that already exist, not to plastic flow.

Contrary to a widespread belief, deep-focus earthquakes do not necessarily require strength for an accumulation of stresses, but they do require either large strength or a high coefficient of viscosity (see Chap. XI).

The problem of the breaking strength is still more complicated.[54] As has been mentioned, it is frequently smaller than the strength for small confining pressures but increases faster with the increase of the confining pressure. Otherwise, near the surface of the earth the breaking strength seems to be of the same order of magnitude as the strength. According to Greenwald[40] the crushing strength of coal was found in laboratory tests to be between 1×10^8 and $3\frac{1}{2} \times 10^8$ dynes per square centimeter, whereas the lateral bearing strength of the Pittsburgh coal bed might be in the neighborhood of 3×10^8 dynes per square centimeter when large areas were involved.

According to Griggs,[15] a rock is more ductile when the differential pressure is increased rapidly than when the pressure is slowly increased. On the other hand, "when a rock enters the region of plastic flow, it will not deform indefinitely, but will rupture if the deformation is carried far enough" (Ref. 15, page 576). Griggs has defined the *fundamental strength* as the differential pressure that a body is able to withstand under given conditions of temperature and confining pressure without rupturing or deforming continuously. It is given by the smallest value of the different types of strength and determines whether a body can stand a given stress without breaking or flowing.

PLASTIC FLOW

As soon as the stress difference in a body increases beyond its strength, plastic flow begins. Again, no equations for this process have been derived from purely theoretical reasoning, but the fundamental equation that was suggested first by Maxwell[20] is considered a fair approximation. Just as we have neglected certain terms in Eq. (100) to get a useful equation (103) for the internal friction, we suppose now that the strain S in a solid undergoing plastic flow is given by

$$dt. \qquad (108)$$

μ is the coefficient of rigidity, F the tangential stress, t the time and τ a new quantity which changes depending on the conditions but which is assumed to be constant to avoid too great theoretical complications. τ controls the time required by a given change due to plastic flow. Usually the quantity $\nu = \mu\tau$ is called *coefficient of viscosity*. τ has the dimension of time, ν is usually given in poises

(1 poise = 1 dyne sec./cm.^{2} = 1 gr./cm. sec.).

The first term of Eq. (108) gives the effect of the elastic change; the second gives the effect of the plastic flow. In many instances a better agreement between observations and calculated values is obtained by adding another term in Eq. (108) (see, *e.g.*, Ref. 46).

As in the case of internal friction, the effect of plastic flow upon a process can be found to a first approximation by replacing the terms involving shear products of the type $\mu f(t)$ by[5]

$$e^{-t/\tau}\int_0^t dt\, e^{-t/\tau}\,\frac{\partial f(t)}{\partial t} \tag{109}$$

and products of the type $\lambda f(t)$ by

$$\lambda f(t) + \frac{2}{3}\mu\left[f(t) - e^{-t/\tau}\int_0^t dt\, e^{-t/\tau}\,\frac{\partial f(t)}{\partial t}\right]. \tag{110}$$

In problems of plastic flow as well as in problems of internal friction, terms containing only the bulk modulus (type: $kf(t)$) but neither of Lamé's constants remain unchanged. A process involving only a change in volume but no shear occurs, to a first approximation, always in a purely elastic way without plastic flow.

If the distortion S is kept constant, we find from Eq. (108)

$$\frac{\partial F}{\partial t} = -\frac{1}{\tau}F, \qquad F = F_0\, e^{-t/\tau}, \tag{111}$$

which means that τ is the time in which the stress decreases to $1/e$ by plastic flow, if the distortion is kept constant. τ, therefore, is frequently called the *time of relaxation* due to plastic flow.

The process producing plastic flow in solids differs fundamentally from the process involved in internal friction. This can best be seen by comparing the equations for solids.

Internal friction:

$$\frac{\partial S}{\partial t} = \frac{F}{\eta} - \frac{S}{\vartheta}. \qquad S = \frac{F}{\mu} - \vartheta\,\frac{\partial S}{\partial t}. \tag{112}$$

Plastic flow:

$$\frac{\partial S}{\partial t} = \frac{F}{\nu} + \frac{1}{\mu}\frac{\partial F}{\partial t}. \qquad S = \frac{F}{\mu} + \frac{1}{\nu}\int F\, dt. \tag{113}$$

Purely elastic:

$$\frac{\partial F}{\partial t}. \qquad S = \frac{F}{\mu}. \tag{114}$$

Although the distortion S in a body with internal friction is smaller than in a purely elastic substance, it is larger in a plastic material. As we have seen (Eq. 105), a fluid with internal friction behaves in the way usually described as "viscous." For a fluid having the property called *plasticity*, we find that S tends to infinity: the fluid behaves like an ideal fluid. Thus, in solids as well as in fluids, the two properties are fundamentally different. Viscous flow is made possible in fluids by the stiffening effect of internal friction and in solids by the weakening effect of plasticity, which has been called "elasticoviscosity" by Michelson.[31]

Though for long-period motions the effect of internal friction is small, it is large for a plastic solid; the last term in Eq. (108) is large in this case, and the equation approaches the form $F = \nu(\partial S/\partial t)$, which, as we have seen, is characteristic for a viscous fluid. For short-period movements, however, the last term becomes negligible, and the plastic solid behaves almost like a perfect elastic body. In this case, the effect of internal friction delays the movement so that it approaches the conditions of a perfect rigid body.

For intermediate cases, both internal friction and plasticity may produce simultaneously noticeable deviations from the perfect elastic conditions. Here the equation may be written in the form

$$\mu\left(S + \vartheta \frac{\partial S}{\partial t}\right) = F + \frac{1}{\tau}\int F\,dt. \tag{115}$$

Fortunately, however, the conditions in geophysical problems have so far been such that either plasticity or internal friction has too small an effect to be considered.

For stress differences below the strength, theoretically there is no plastic flow. Up to that point we have $S = F/\mu$ if we neglect the deviation from Hooke's law. If F exceeds the strength, Eq. (108) must be used. Mathematically, both equations can be combined by assuming that ν is infinite if F is below the strength and has a finite value if F is greater than the strength. However, the observations show complications. During plastic flow, the strength seems to increase slightly. Moreover, the rate of flow is influenced by the way in which the stresses change[15,46] and the history of the material (Ref. 46, page 173).

Some authors believe that the maximum stress difference controls plastic flow. Mises[21] considers the sum

$$(a - b)^2 + (b - c)^2 + (c - a)^2$$

as most important for plastic flow, where a, b, and c are the principal stresses. Others, *e.g.*, Prager,[22] consider best a combination of both, one for the beginning of the flow and the other for the later part. In spite of many investigations[16,23,24,25,38,39,46,47] and the importance of these matters for technical purposes, the problems concerning plasticity present many unsolved questions.

The ratio of the coefficient of viscosity divided by the density is called *kinematic viscosity.*

Relatively few observations of the coefficient of viscosity ν and the time of relaxation τ are available [for methods see, for example, Refs. 46 (page 15) and 51]. The following values give an idea of the order of magnitude. ν is given in poises (= grams per centimeter per second), and τ in seconds.

TABLE 67

VISCOSITY ν (IN POISES) AND TIME OF RELAXATION τ (IN SECONDS)
(Order of magnitude)

Viscosity at which gas bubbles can escape[46]	10^2	
Viscosity at which glass can be wound on glass-blower's pipe	10^3	
Viscosity at which glass can be blown[46]	10^7	
Shoemaker's pitch at 50°C.[51]	10^4	10^{-}[illegible]
Shoemaker's pitch at 15°C.[51]	10^8	10
Asphalt at 20°C. (see Table 69)	10^7	
Ice[51]	10^{13}	500
Iceland spar at 18°C.[51]	10^{16}	10^5
Rock salt at 18°C	10^{18}	10^7
Rock salt at 80°C.[51]	10^{17}	10^6
Solenhofen limestone[32]	10^{21}	10^{10}

The coefficient of viscosity depends on various quantities as has been mentioned before. For fluids, the change of ν in time under otherwise unchanged conditions has been investigated by Umstätter.[48] He found the following equation to be a good approximation:

where ν_0, ν_t and ν_∞ are the coefficients of viscosity at the time 0, t and infinity, respectively.

The effects of temperature and pressure have been investigated by Barus[41] in 1893. Many scientists have dealt with these problems since. In most cases they seem to have been unfamiliar with the earlier literature, as has been pointed out by Ewell[47] (see especially page 227 of his paper). For the effect of pressure p on the coefficient

of viscosity, Barus has given the following empiric equation as a rough approximation:

$$\log \log \nu = A + Bp, \tag{116}$$

where ν = coefficient of viscosity, p = confining pressure and A and B are constants. Hyde[42] has found similar changes of the viscosity of oil with increasing pressure. There is less agreement about the change

TABLE 68

VISCOSITY ν IN DYNES/SQ. CM. OF SEVERAL TYPES OF GLASS AS A FUNCTION OF TEMPERATURE t (°C.)

(A. After English.[44] B. After Washburn and Shelton[45])

A

t	ν	t	ν	t	ν
1310	77	1410	70	1400	51
1240	96	1315	161	1310	83
1136	380	1194	547	1205	243
1028	1.8×10^3	1100	1.9×10^3	1014	2.7×10^3
920	1.1×10^4	992	8.9×10^3	906	1.6×10^4
790	2.7×10^5	650	3.2×10^8	745	1.6×10^6
600	7.3×10^8	555	9.5×10^{10}	600	2.1×10^9
485	5.6×10^{12}	505	6.2×10^{12}	500	3.15×10^{12}

B

t	70 % SiO_2 30 % Na_2O 0 % CaO	82.6 % 17.4 % 0 %	70 % SiO_2 20 % Na_2O 10 % CaO	73.5 % 16.5 % 10.0 %
1500	38	214	38	48
1400	77	434	87	150
1300	160	1200	214	500
1200	420	4900	600	1.8×10^3
1100	1.3×10^3	3.2×10^4	2×10^3	7.1×10^3
1000	4.4×10^3	2.5×10^5	8.5×10^3	2.9×10^4
900	2×10^4	?	4.8×10^3	1.2×10^5
800	1.4×10^6	?	4.3×10^5	5.4×10^5
750	4.3×10^5	?	?	?

of viscosity with temperature. Whereas Barus considered the following relation as a fair approximation:

$$\log \nu = C - Dt, \tag{117}$$

where t = temperature, and C and D are constants, Le Châtelier[43] preferred the equation

$$\log \log \nu = E - Ft, \tag{118}$$

which agrees better with observations on the viscosity of glasses, as given in Table 68.

For asphalts, Pittman and Traxler[50] found the values given in extenso in Table 69. A summary for other substances has been given by Houwink (Ref. 46, page 144).

TABLE 69
VISCOSITY ν FOR ASPHALTS AS A FUNCTION OF TEMPERATURE t
(After Pittman and Traxler[50])

t, °C.	Viscosity in various asphalts, poises		
15	9.2×10^7	6.5×10^7	3.7×10^7
25	7.0×10^6	3.3×10^6	3.7×10^6
40	2.3×10^5	8.1×10^4	1.3×10^5
60	8.7×10^3	2.7×10^3	4.3×10^3
80	7.0×10^2	2.5×10^2	3.7×10^2
100	1.0×10^2	3.3×10	6.3×10
130	1.1×10	4.8	8.7

Very complicated changes in viscosity have been found for molten sulphur (see, *e.g.*, Ref. 46, page 364). With the gradual heating of a sulphur melt from 150°C. the viscosity increases rapidly to a maximum at 187°C., which is about 6,000 times the value at 150°C., and then decreases gradually upon further heating. At 400°C. it is still about ten times as high as at 150°.

By using a new apparatus for measuring the deformation of rocks when differential stresses below the elastic limit are applied for long time intervals (months or years), Griggs[32] has found that Solenhofen limestone has an "exponentially decreasing strain rate which after three months time is apparently approaching a constant flow rate. This corresponds to a viscous flow with a viscosity of 1.8×10^{21}

If we investigate plastic flow in the interior of the earth, we must consider that its mechanism there may be different at different depths. The best we can do is to assume that Eq. (108) gives a sufficient approximation for the process. It is not unlikely that the coefficient of viscosity found for the various depths is a combination of different quantities of which one dominates in one range of depth and others at deeper or shallower levels. Creep as defined before possibly plays a more important role at larger depths but would be included in the "coefficient of viscosity" as long as we cannot discriminate between

macroscopic and microscopic flow at large depths. Griggs's results in regard to Solenhofen limestone, mentioned above, indicate that possibly various processes are involved at relatively low pressures.

The first estimate of the coefficient of viscosity inside the earth seems to have been made by Schweydar.[26] He found that the observed body tides of the earth would be impossible, if there were a layer 100 km. thick with a viscosity of 10^9 poises (more than that for pitch at normal temperature) or a layer with a thickness of 600 km. and a viscosity of 10^{13} poises (about as viscous as ice).

The best data available have been derived from the uplift in areas of Pleistocene glaciation, especially in Scandinavia. From the observed gravity anomalies A, the rate of rise v, the diameter D of the rising area and the depth H to which the currents extend that are connected with the rise at the surface, Vening Meinesz[27] calculated the coefficient of viscosity in the upper half of the mantle by using the equation

$$\nu = \frac{g\,D^2 A}{16\pi\kappa H v}, \tag{119}$$

where κ = Newtonian gravitation constant. He used the following values: $D = 1{,}000$ km., $A = 40$ milligals, $H = 1{,}200$ km., $v = 3.2 \times 10^{-8}$ cm. per second (1 cm. per year) from present observations and found $\nu = 4 \times 10^{22}$ poises. If the general assumptions on the mechanism of the process are correct, the principal possible errors are in A, which may be affected by local gravity anomalies from other sources, and in H. If the currents extend down to the core, H would be about twice as large: however, it could be very much smaller than has been assumed by Vening Meinesz. In this case the value of ν would increase. As A cannot be too small, but rather too large, we find that the value of viscosity found in this way is probably correct within one order of magnitude.

Haskell[28] proceeded in a different way. He gave a formal solution for the motion of a highly viscous fluid when a symmetrical pressure is applied at the surface. This he applied to the subsidence of a cylindric body of constant thickness and to the recovery after the removal of the load. The resulting equations Haskell applied to the plastic recoil of the earth after the disappearance of the Pleistocene ice in Fennoscandia. Using the changes of the surface during the last 7,000 years as given in the following tabulation, he found the values of the viscosity given in the last column (supposing that the density is 3):

Year a of beginning for the calculation	Uplift, cm./year	Deviation of surface from equilibrium, m.	Viscosity, poises $\times 10^{-22}$
5,000 B.C.	3.9	147	0.8
4,000 B.C.	2.7	118	1.0
3,000 B.C.	2.2	94	0.9
2,000 B.C.	1.8	74	0.9

which gives a coefficient of viscosity of the order of 10^{22} poises. In connection with the assumptions of Haskell, Vening Meinesz[29] has pointed out that the deviations assumed by Haskell refer to the surface today but that this is still about 180 m. lower than its position of equilibrium. By adding these 180 m. to Haskell's values as presented above, he found 3×10^{22} poises for ν by using Haskell's as well as a somewhat simplified method.

The agreement between the value found in differing ways in the applied theory and in the quantities on which they are based makes it probable that the order of magnitude of 10^{22} for the coefficient of viscosity in the upper half of the mantle of the earth is correct. However, we must keep in mind that the foregoing has the same meaning as the statement that the earth as a whole is about as rigid as steel. The values for different depths may vary by several powers of 10. Vening Meinesz[29] has made an attempt to find the change of viscosity with depth. He found a decrease with depth from the order of 10^{23} poises at a depth of about 200 km. to about 3×10^{22} poises near the core of the earth. However, this can be considered only as a first attempt. As ν is proportional to the rigidity, it decreases suddenly by at least a few orders of magnitude at the surface of the core and approaches the internal friction, if the core is fluid. Supposing that this is correct, Jeffreys[35] has found that the small damping of longitudinal waves that have passed the core indicates a viscosity of the order of 10^{10} poises or less inside the core.

The value of τ, which corresponds to $\nu = 10^{22}$ poises in the mantle of the earth, is of the order of 10^{10} sec., or of several hundred years.

The importance of plastic flow for tectonic processes and mining operations has been pointed out by Nádai[24] and others.[30,40] The theory of the effect of plastic flow on geophysical processes has been studied by Jeffreys (Ref. 4, pages 265–267). We can now apply his equations to the new findings. With $\tau = 10^{10}$ sec., the effect on earthquake waves is practically zero, so that this assumption in calculating the coefficient of internal friction from the decrease of the amplitude of earthquake waves was correct. The damping of the polar move-

ment with Chandler's period is given by a factor $e^{-t/3\tau}$. With $\tau = 10^{10}$ sec., plastic flow would decrease the amplitude of the polar movement to $1/e$ in about 1,000 years.

The lag in phase of the body tides of the earth is given by $T/2\pi\tau$. T is their period of one-half lunar day. As τ is of the order of 10^5 days, the phase lag is practically zero, in agreement with observations. Theoretically, the phase lag would exceed a few degrees only if ν would be of lesser order of magnitude than 10^{17} poises, or τ below 10^5 sec.

SUMMARY

In addition to the elastic processes, in which, corresponding to the theory, we assume a complete sudden distortion with the action of a stress difference and immediate complete return to the original form when the stress difference returns to zero, there are other processes that do not follow this supposition. Elastic afterworking, due perhaps to imperfections of crystal structures, internal friction and possibly other causes, delay the elastic processes. On the other hand, creep, probably due to movements along planes of larger dimensions than those of crystals, produces permanent changes which do not permit the return to the original form after the stresses have ceased. If a certain minimum stress difference, the strength, is reached, plastic flow begins, unless the breaking strength of the body is smaller and has previously allowed rupture. If the stress differences accumulate faster than they are reduced by plastic flow, rupture will occur at a higher stress difference regardless of plastic flow, especially in the case of a large coefficient of viscosity.

The theory of these processes is not yet well understood. Possibly two or more different processes are involved. Observations of the quantities in laboratories are scanty, and the data for the earth concern only the upper half of the mantle. Moreover, such data are uncertain by one order of magnitude or more.

The strength is well-known only for the surface layers, where it is of the order of magnitude of 10^9 dynes per square centimeter. It probably does not change much in the upper 30 or 40 km. and probably decreases to 10^7 dynes per square centimeter or less below that depth, since otherwise isostasy would be difficult to explain. The internal friction seems to be of the order of 10^9 poises near the surface and possibly increases slightly with depth. It is probably small in the core. The upper half of the mantle behaves as if it had a coefficient of viscosity of the order of magnitude of 10^{22} poises throughout. It is probably much smaller in the core. Little is known for rocks at the

surface, where a value of the order of magnitude of 10^{18} has been found for rock salt and of 10^{21} for Solenhofen limestone. However, in this case, the coefficient of viscosity changed with time, and the experiments indicate that the processes involved are expressed only very roughly by the equations. This is true for all the processes which have been discussed in this chapter.

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CHAPTER XVI

SUMMARY*

B. Gutenberg

(*Professor of Geophysics, California Institute of Technology, Pasadena*)

About two or three thousand million years ago, the earth formed as a separate body. During at least a part of its early history, it was liquid, and in this state its free iron together with materials soluble in it settled to the center, thus forming the core of the earth with a radius at present of about 3,500 km. and the silicate shell which has separated into various layers. A well-marked change occurs at a depth of about 1,000 km. Though above this depth the velocity of elastic waves increases quite rapidly with increasing depth, the increase is smaller and more irregular below this depth. This was first recognized by E. Wiechert. Except for the Pacific basin and possibly the Arctic basin, a sudden change in material has been found at depths ranging from about 20 in the Atlantic and Indian oceans to about 40 km. or slightly more in the central parts of the continents; there are additional discontinuities within the upper part of the continental layer. Several authors have suggested first-order discontinuities in the mantle at various depths; however, none of these suggestions has been generally accepted. At a depth of between 200 and 500 km. the velocity of earthquake waves increases relatively fast; whether this includes a discontinuity is still an open question. The additional findings that no earthquakes seem to originate deeper than about 700 km. below the surface and that at about the same depth the electric conductivity seems to increase rapidly may suggest a change in properties at a depth of order 500 to 700 km. below the earth's surface.

The dominating material of the upper sublayer in the continental lithosphere is granite. The fact that enormous masses of basalt of a composition nearly uniform in space and time have been erupted led to the conception of a world-encircling shell of basalt at no great depth. No large bodies of tachylite or of eclogites with the chemical composi-

* The author of this summary is alone responsible for statements as to controversial points contained in it, which do not always coincide with the opinions of other members of the committee.

tion of basalt have been found, and there is no field evidence for dunite or peridotite.

The average of analyses of igneous rocks contains about 69 per cent SiO_2; about 15 per cent Al_2O_3; about 5 per cent CaO; between 3 and 4 per cent of Fe_2O_3, FeO, MgO, Na_2O, K_2O each; about 1 per cent of H_2O and TiO_2; these oxides together constitute over 99 per cent of the actual eruptive material of the globe. The four most abundant elements in the igneous rocks that have been studied are oxygen (47 per cent), silicon (28 per cent), aluminum (8 per cent) and iron (5 per cent). For the whole earth, iron seems to take first place with probably 40 ± 10 per cent; the oxygen content is estimated to be between 20 and 30 per cent, silicon to be between 10 and 15 per cent, and all others to be less than 10 per cent, with only Mg, Ca, Ni and possibly Al with more than 1 per cent. The relative abundance of the various elements in the earth seems to be of the same order as in the sun, except for the light gases which are much more abundant in the sun. On the other hand, the composition of the earth corresponds roughly to that of 55 parts of stone meteorites plus 40 parts of iron meteorites plus 5 parts of troilite.

During the early youth of the earth, cooling was very great, and crystallization of the light surface layers probably began less than 100,000 years after the origin of the earth as a separate body. Because of the relatively large difference in density between the crustal layers, the crystalline blocks forming at the surface could not sink very deep, and relatively soon afterward the earth was covered by a solid crust, which, however, may have been partly destroyed and rebuilt many times during the early history of the earth. Nevertheless, this crust was sufficient to reduce the speed of cooling considerably, since this was then controlled by thermal conduction in the crust, while earlier several systems of convection currents may have speeded up the cooling. In the deeper layers, such convection currents probably persisted in some parts of the earth and may still exist, moving, however, at small velocity.

The amount of cooling was and is greatly reduced by the heat developed in radioactive processes. Most of the heat passing outward through the surface of the earth arises from the decomposition of radioactive matter. If the estimates of geochemists on the amount of radioactive matter inside the earth were correct, heat generated in this way would exceed that being carried through the surface by pure conduction. This hypothesis has suggested the idea that the excess of heat has been dissipated through the action of subcrustal currents with mountain making as a by-product. On the other hand,

most geophysicists hold the opinion that the earth is cooling and that the amount of heat produced inside the earth is smaller than that calculated by assuming that radioactive matter exists in amounts as great as most geochemists believe. The amount of this heat must decrease in the course of time, since the less stable elements, which produce the greater part of it, get scarcer. For this reason, the cooling of the earth must speed up. Geologic evidence indicates that the earth's crust has not been cooling steadily.

In general the temperature increases with depth in the earth's crust by between 15 and 35°C., probably by about an average of 20°C. per kilometer. The temperature of lavas in volcanoes is in general 1100 ± 100°C. Beginning at a depth of about 50 km., the increase in temperature probably becomes noticeably less, and the temperature in the core is probably of the order of 2000 or 3000°C. As the increase of melting point with pressure is not accurately known, no conclusions can be drawn as to the depth where the melting point is reached for a given material. It is not unlikely that this point occurs in the mantle at a depth of about 60 km., where the elastic constants seem to decrease slightly. There may even be other regions at greater depths where the temperature is below the melting point, and this is one possible explanation for the irregularities found for the velocities of elastic waves in the deep parts of the mantle.

The density increases from about $2\frac{3}{4}$ near the surface to about $4\frac{1}{2}$ at a depth of 1,000 km., is probably between 6 and 8 at the boundary of the core and may reach or even surpass 12 in the center of the earth. The pressure of 1 million atmospheres occurs at a depth slightly greater than 2,000 km., of 2 millions at about 3,500 km., and reaches about $3\frac{1}{2}$ million atmospheres in the center of the earth. Gravity does not change much in the mantle, where, as a whole, it increases slightly. In the core it decreases to zero at the center.

The bulk modulus of the crust is of the order of magnitude 10^{12} dynes per square centimeter and increases to the order of 10^{13} in the deep parts of the core. The rigidity is of the order of 5×10^{11} in the crust and increases to about 4×10^{12} dynes per square centimeter approaching the core. Inside the core, it seems to be much smaller, but even its order of magnitude is unknown there; the assumption that the core is fluid does not lead to contradictions with the known facts.

The strength (resistance against plastic flow) is of the order of 10^9 dynes per square centimeter near the surface. Its value at greater depths is not known; the small differences in gravity over the whole earth, regardless of mountains or oceans (isostasy), seem to indicate

that below the crust plastic flow occurs even with small differences in stresses and that strength there is at least a few orders of magnitude smaller than at the surface. However, the discussion of this subject is hampered by the lack of precise meaning to be attached to such terms as *strength, viscosity, internal friction* and *creep*. If we consider isostatic readjustment as due to plastic flow, the coefficient of viscosity for the outer half of the mantle as a whole is of the order of 10^{22} poises (1 poise = 1 g. per centimeter second); the observations of the body tides indicate a viscosity of at least 10^{18} poises, and from the polar movements a viscosity of at least 10^{20} poises has been calculated, so that 10^{22} poises seems to be at least a good approximation. This would mean that it takes times of the order of 100 years or more to produce an appreciable plastic flow; for processes of a shorter duration the effect would be similar to that of a very large strength. This viscosity is combined with an internal friction of the order of 10^9 poises for movements with periods over one sec., which produces effects to be considered in investigations on the interior of the earth only for relatively fast movements, such as occur in earthquake waves, but not for polar or tidal movements. For very fast movements the internal friction seems to decrease and seems to be of the order of magnitude of 10^7 poises for waves with a period of 0.01 second.

Although the properties of the interior of the earth are partly known to a first approximation, the major forces producing changes are still being assumed quite differently by various authors. As there is no general agreement even as to whether or not more heat is produced inside the earth than escapes, it is not possible to estimate the contraction due to a probable cooling. Though many still believe that this contraction is sufficient to account for the formation of mountain ranges during the history of the earth, the opinion is held by an increasing number of geologists and geophysicists that the thermal contraction, at best, plays an important but not the dominant role in the tectonic history of the earth. The existence of subcrustal currents is assumed by an increasing number of writers. As to the cause of such currents, opinions differ widely; magmatic differentiation and radioactive heat have been suggested as among the more important causes. The differences in temperature and other properties, between ocean bottom and continent, as well as sedimentation, are regarded as less important. Formation of mountain ranges and extended but narrow deeps, drifting of the continental block or large parts of it with corresponding changes in climate, spreading of the continents are among the deduced consequences of such subcrustal currents. The inequalities of the earth's surface layers add to the stresses produced by the

currents at greater depths. Most of the special hypotheses regarding the development of the earth's crust, however, have bases so uncertain that new suggestions are being proposed at a rapid rate. The actual changes are probably produced and controlled by a complicated combination of processes of the types just mentioned.

Proof that stresses are acting at depths as large as 700 km. is furnished by the deep-focus earthquakes. All observations indicate that these are due to shearing stresses and rupture similar to that in normal shocks. The deepest foci, at depths between about 250 and 700 km., appear to be associated with the boundary of the Pacific basin. Together with facts derived from the study of normal shocks and the geologic evidence, they emphasize the unique structure of the Pacific basin, already mentioned. The intermediate shocks (depths between 60 and 250 km., approximately) are chiefly associated with lines of the tectonic activity that occurred in the vicinity of Tertiary time, whereas the present seismic activity, which frequently occurs along relatively narrow strips of large negative gravity anomalies, indicates the regions where tectonic movements are now going on.

APPENDIX

FREQUENTLY USED CONSTANTS

In calculations concerning the interior of the earth, astronomical and geodetic constants are frequently needed. For this reason some of them are given here:

Semi-axes of the earth's ellipsoid (internationally adopted values):

$$a = 6.378388 \times 10^8 \text{ cm.}$$
$$c = 6.356912 \times 10^8 \text{ cm.}$$

$$\text{Flattening} = \frac{a - c}{a} = \frac{1}{297} = 0.00337.$$

Quadrant of a meridian = 10,002.288 km.; of the Equator = 10,019.148 km.
Surface of the earth = 5.101×10^{18} sq. cm.; volume = 1.083×10^{27} cc.
Average density = 5.52; mass = 5.98×10^{27}g. = 3.00×10^{-6} of the mass of the sun.
Gravity at sea level in latitude φ (internationally adopted):
$g = 978.049(1 - 0.0052884 \sin^2 \varphi - 0.0000059 \sin^2 2\varphi)$ gal.
Especially, at the Equator $g_e = 978.049$; at latitude 45° $g_{45} = 980.6294$; at pole $g_p = 983.221$ gal.
If A and C are the moments of inertia about the axes a and c, respectively, then

Angular velocity of the earth's rotation, $\omega = 15.04106863''$ per second = $7.292115851 \times 10^{-5}$ radians per second; velocity of a point on the Equator = 46,500 cm. per second.
Day (solar) = 86,400 sec.; sidereal day = 86,164.09 sec.
Centrifugal force at the Equator divided by gravity at the Equator =

Mean distance earth to sun = 149.5×10^6 km. = $23,439 \times$ radius of the earth Eccentricity of orbit = 0.01675.
Average daily movement around the sun = 3,548.19″; average velocity = 29.8 km. per second.
Sidereal year = 365.256 (solar) days = 3.1558×10^7 sec.
Tropical year = 365.2422 days.
Period of precession of the equinoxes = 25,735 years.
Mean distance moon-to-earth = 384,400 km.; eccentricity of moon's orbit = 0.0549; density of moon = 3.335; mean radius = 1,738 km.; mass = 0.012265 of mass of earth.
Period of sidereal revolution of moon = 27.32166 days = 2.3606×10^6 sec.

UNITS

Throughout the book the use of the c.g.s. system is preferred. Below are given some equivalents of units which are used in problems concerning the interior of the earth.

Linear Measures

1 cm. = 0.39370 in. = 0.032808 ft.
1 km. = 10^5 cm. = 0.62137 mile.
1 fathom = 6 ft. = 1.8288 m.
1 nautical mile = 1.85325 km. = 1 min. of arc on the earth (Clarke's spheroid of 1866).

1 in. = 2.54001 cm.
1 ft. = 30.480 cm.
1 mile = 1.60935 km.

Square Measures

1 sq. cm. = 0.1550 sq. in. = 0.0010764 sq. ft.
1 sq. km. = 10^{10} sq. cm. = 0.3861 sq. mile.

1 sq. in. = 6.452 sq. cm.
1 sq. ft. = 929.0 sq. cm.
1 sq. mile = 2.5900 sq. km.

Cubic Measures

1 cc. = 0.0610 cu. in. = 0.000035314 cu. ft.

1 cu. in. = 16.387 cc.
1 cu. ft. = 28317 cc.

Mass

1 kg. = 10^3 g. = 2.20462 lb.

1 lb. = 0.453592 kg. = 453.592 g.

Velocity

1 ft. per second = 30.480 cm. per second = 1.0973 km. per hour.
1 mile per hour = 1.6093 km. per hour = 44.704 cm. per second.
1 cm. per second = 3.728×10^{-4} mile per minute = 0.02237 mile per hour.
1 km. per hour = 27.7778 cm. per second.

Acceleration

1 gal = 1 cm. per second per second = 0.032808 ft. per second per second.
1 milligal = 10^{-3} gal.
g = 980.665 gal. = 32.174 ft. per square second.
(International standard.) (For details see preceding table.)

Force

As units of force, frequently *gram weight*, *kilogram weight*, *pound weight*, etc., are used. We denote them here by g.*, kg.*, lb.*, etc.
1 dyne = 0.0010197 g.* = 2.2481×10^{-6} lb.*
1 megadyne = 10^6 dynes.
1 g.* = 980.665 dynes.
1 lb.* = 4.4482×10^5 dynes.

Pressure

The c.g.s. unit of pressure is the *barye;* there is besides by international agreement the *bar* which is 10^6 baryes. Although the bar is used internationally in

meteorology as here defined, physicists use frequently, especially in Europe, the word bar as synonymous with barye as c.g.s. unit. In many handbooks the barye is not mentioned. Thus, in using data given in bars as unit, one must make sure first which of the two units is used. The same is true for the *megabar*, which may mean 10^6 bars of either unit bar. The following definitions are based on the international agreement.

1 barye = 1 dyne per square centimeter = 9.8692×10^{-7} atmospheres = 10^{-6} bar = 1.4504×10^{-5} lb.* per square inch.
1 bar = 10^6 baryes = 10^6 dynes per square centimeter = 0.98692 atmosphere = 14.504 lb.* per square inch = pressure of 750.06 mm. of Hg.
1 millibar = 10^{-3} bar = pressure of 0.75006 mm. Hg.
1 megabar = 10^6 bars.
1 kg.* per square centimeter = 0.980665×10^6 dynes per square centimeter = 14.233 lb.* per square inch.
1 lb.* per square foot = 4.7254×10^{-4} atmosphere = 478.80 dynes per square centimeter.
1 lb.* per square inch = 0.068046 atmosphere = 0.068947 bar = 6.8947×10^4 dynes per square centimeter.
1 atmosphere = 1.0332 kg.* per square centimeter = 1.01325 bars = 14.7 lb.* per square inch = pressure of 760 mm. Hg.

Temperature and Temperature Gradient

5°C. = 9°F.
Temperature a°C. = $(\frac{9}{5}a + 32)$°F.
1°F. per foot = 0.0182269°C. per centimeter.
1°C. per centimeter = 54.864°F. per foot.
1 ft. per degree Fahrenheit = 0.54864 m. per degree centigrade.
1 m. per degree centigrade = 1.82269 ft. per degree Fahrenheit.

Miscellaneous

1 erg = 1 dyne centimeter.
1 joule = 10^7 ergs = 0.102 m. kg.* = 0.737 ft.-lb.
1 watt = 10^7 ergs per second = 1 joule per second.
1 poise = 1 g. per centimeter per second.
1 rhe = 1 cm. sec. per g.
1 gauss (1Γ) = electromagnetic unit = 1 g.$^{\frac{1}{2}}$ cm.$^{-\frac{1}{2}}$ sec.$^{-1}$
1 gamma (γ) = 10^{-5}Γ.
1 radian = 0.159155 circumference = 57.29578° = 57° 17′ 44.8″ = 2.06265×10^{5}″.
1″ = 4.84814×10^{-6} radians.
sin 1″ = 4.84814×10^{-6}.

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